GEOTHERMAL INVESTIGATIONS IN IDAHO

EVALUATION OF THE BOISE GEOTHERMAL SYSTEM

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EVALUATION OF THE BOISE GEOTHERMAL SYSTEM

FINAL REPORT TO

IDAHO DEPARTMENT OF WATER RESOURCES

BOISE, IDAHO

in fulfillment of

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DISCLAIMER

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ABSTRACT

Evaluation of the Boise Geothermal System is presented in two parts. Part I discusses the physical setting of the hydrogeologic system and Part II reviews the geochemical data.

The geothermal aquifer is a fractured media groundwater system which produces hot artesian water from the network of fractures of the Boise Frontal fault system and the fractured, layered rhyolites and interbedded sediments of the Idavada Group, and from fracture zones within the Idaho batholith. Prior to 1982-83, the system appears to have been at or near equilibrium. However, increases in production since that time have resulted in a general decline in recovery levels, and a pattern of evidence indicating interconnection between the Capitol Mall-Reserve Park (BGL) portion of the aquifer system and the Warm Springs portion is emerging.

In the Boise geothermal system, subsurface residence times derived from radiocarbon activity, range from 6700 to 17,000 years in four analyses. Boise Warm Springs Water District (BWSWD) well water has the greatest carbon -14 residence time, the greatest temperature of the studied wells, and the lowest $\delta^{13}C$ compared with the other wells of the Boise system. The Capitol Mall No. 2 (CM No.2) well has the shortest residence time, the lowest pH, and the lowest fluorite solubility product. Data suggest that the BWSWD water is the least mixed sample and is probably derived mostly from the deep circulation system, and that other wells draw some of their water from systems of

shallower circulation. It is recommended that future geochemical study be directed toward quantifying the mixing of deeply circulating water with the waters that have only circulated in the shallow stratified rocks. Future analysis should focus upon constituents or properties of the deeply circulating water that might be conservative in the mixing process. Temperature, fluorite solubility products, and lithium are suggested. Such a study could help to quantify the subsurface discharge of the deeply circulated water into the stratified rocks. Knowledge of the amount and the region of that discharge may help in design of future reinjection and production wells.

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SUMMARY

The Boise Geothermal system is a fractured-media groundwater system which produces hot artesian water from the fault zone along the Boise front and from subsurface rhyolite layers and interbedded sediments within the Idavada Group. The aquifer is confined by clayey, basaltic, volcaniclastic sediments and tuffs of low permeability. The wells of two of the three principal producers of hot water, Boise Warm Springs Water District (BWSWD) and Boise Geothermal Limited (BGL), produce from a fracture network along the Boise Front. The third, the State of Idaho Capitol Mall system, produces primarily from Idavada Group rhyolites approximately 2500 feet southwest of the Boise Front. Only the Capitol Mall Project reinjects its production into the aquifer. Temperatures of the water upon reinjection are commonly 15° to 30°F cooler than when produced. Water from BGL wells after its use for heating, is gathered and discharged into the Boise River. BWSWD production is used for heating and as domestic hot water supplies. After use, the geothermal water is discharged into canals in the vicinity or into the city sewage system.

BWSWD wells have been in use since about 1892; however, records of their production prior to June 1977 are not available. Since 1977 partial records on water levels and production have been accumulated by the District. The District records indicate that their production has ranged from approximately 300 million gallons to 235 million gallons per year for the period 1978

through 1986. The principal control on production at BWSWD is demand for heating. However, in recent years, 1984 through 1986, large drawdowns to the level of the pumps during the peak heating season have caused cavitation and limited their production.

Boise Geothermal Limited began production in October of 1983. In the subsequent three years 1983-84, 1984-85, and 1985-86, the BGL wells have produced approximately 166.7, 121.4, and 176.8 million gallons respectively. Capitol Mall production began in 1982 and has produced an average of approximately 190 million gallons per year in the subsequent four years including 1982-83.

This increased development and use of the system since 1982 and 1983 has been attended by an annual decline in the level to which the piezometric surface recovers during the summer period of low pumpage. Between September 1983 and September 1986, maximum recovery levels in observation wells monitored have declined an average of 5 feet per year. During the same period, the maximum recovery levels in the BWSWD pumping wells have declined an average of 8 feet per year.

Information on water levels in the BLM observation well prior to 1983 are scanty; however, the data available indicate that although there were annual water-level fluctuations ranging through 8-12 feet, maximum recovery levels peaked at elevations near 2760 feet in 1978, 1981, and 1982. In addition, according to BWSWD data, surface flows consistently occurred at BWSWD wells 1 and 2 during maximum recovery in the years 1978 through 1983.

These data allow the suggestion that prior to 1983-84 the BWSWD and BLM-BGL-CM portions of the aquifer were near equilibrium.

The data do not indicate a concomitant expansion of production by BWSWD after 1983. The most likely explanation for the increased declines in both portions of the aquifer system since 1983 are withdrawals by BGL which began in the fall of 1983, and to a lesser extent Capitol Mall which began withdrawals but also began reinjection in 1982. This interpretation requires that some degree of interconnection exists between the BLM-BGL-CM and BWSWD portions of the aquifer.

Geologically and hydraulically such interconnection is likely. Review and interpretation of the geologic and hydrologic data currently available allows suggestion of the following working aquifer model which will require testing, revision, and ultimately confirmation or rejection. The aquifer consists of interlayered sequences of rhyolites and sediments, and granites which have been divided into major "blocks", sections, or segments by relatively large fracture zones. Portions of the fracture zones separating the segments apparently range from nearly open conduits such as that which BWSWD wells 1 and 2 and the BGL wells encountered, to portions which are partially or nearly filled with secondary deposits including zeolites, clay, silica, etc. The open fracture zones have the largest transmissivities (highest permeabilities); whereas, partially cemented fracture zones and the interfracture zone "blocks" or segments have a wide range of transmissivities (permeabilities).

The broad variation in values of transmissivity that have been estimated by pump tests and long term withdrawals from the aquifer probably reflect this segmentation of the aquifer.

Although the aquifer seems to have been near or at equilibrium prior to the 1983-84, the current data base does not yet allow a prediction as to where or if a new equilibrium level will be established. It is interesting to note that the total annual production for BWSWD and BGL combined for 1983-84, 1984-85, 1985-86, and 1986-87 are 459.6, 380.4, 436.2, and 390 million gallons respectively. The total combined withdrawals by these two producers have generally declined since 1983-84. Yet, water levels within those portions of the system are declining at increasing rates with even greater increases in the rates evident from 1986-1987 data. In addition to the decline in production since 1983-84 the balance of production has shifted. In 1983-84, 63.7% of the withdrawals were by BWSWD. In 1986-87 BGL will produce about 51% of the total. In addition to the BWSWD, BGL, and CM withdrawals, an estimated 15-25 million gallons of geothermal water is being produced annually for irrigation by the Boise City Parks Quarry View Well, and by the Idaho Botanical Garden from the State Well. These withdrawals also commenced about 1983-1984.

If as planned, even greater production from the system occurs, water levels within the system can be expected to decline more rapidly and a new piezometric equilibrium level delayed. The increasing rates of decline in the recovery levels evident since 1983 occurred without a corresponding increase in geothermal fluid production by the principal producers.

Development of the geothermal aquifer is currently out-stripping knowledge and understanding of the aquifer; thus, ability to prudently manage the system is hampered. It is clear, however, that current production from the aquifer in the Boise vicinity exceeds recovery. Serious consideration should be given to requiring reinjection of produced waters where feasible. Although there are insufficient data to predict accurately the effects of reinjection, current interpretation of the aquifer system and the apparent degree of interconnection that exists within it, suggests that an effective reinjection program will significantly affect the economic productivity and extend the life of the resource.

INTRODUCTION

Warm geothermal water produced from three well fields (Fig. 1) is currently used to heat about 2,000,000 square feet of office building and residential housing in downtown Boise. The Boise Warm Springs Water District (BWSWD) No. 1 and 2 wells have been in use since 1892 and have typically produced between 235 and 300 million gallons/year of approximately 165°F water for space heating and domestic use. The State of Idaho Capitol Mall No. 2 well has produced water at an average rate of about 190 million gallons/year and temperature of 157°F since October, 1982. The Boise Geothermal, Ltd. wells have produced 166.7, 121.4 and 176.8 million gallons of 165°F water annually in the three years since October 1983. Demand for the hot water generally increases from the beginning of the heating season in September to a peak in January, then begins to decline again through May. In response to these withdrawals, water levels within the aquifer system currently decline from their peak recovery in September to their lowest levels in late February and early March, at which time the recovery portion of the cycle begins. Approximately 30 days lag time exists between the peak in the production cycle and the greatest drawdown created by the production. Of the three major producers from the system, only the Capitol Mall (CM) reinjects its geothermal water. After heating use, Boise Geothermal Limited regathers its effluent into a single pipeline and discharges it into the Boise River. BWSWD

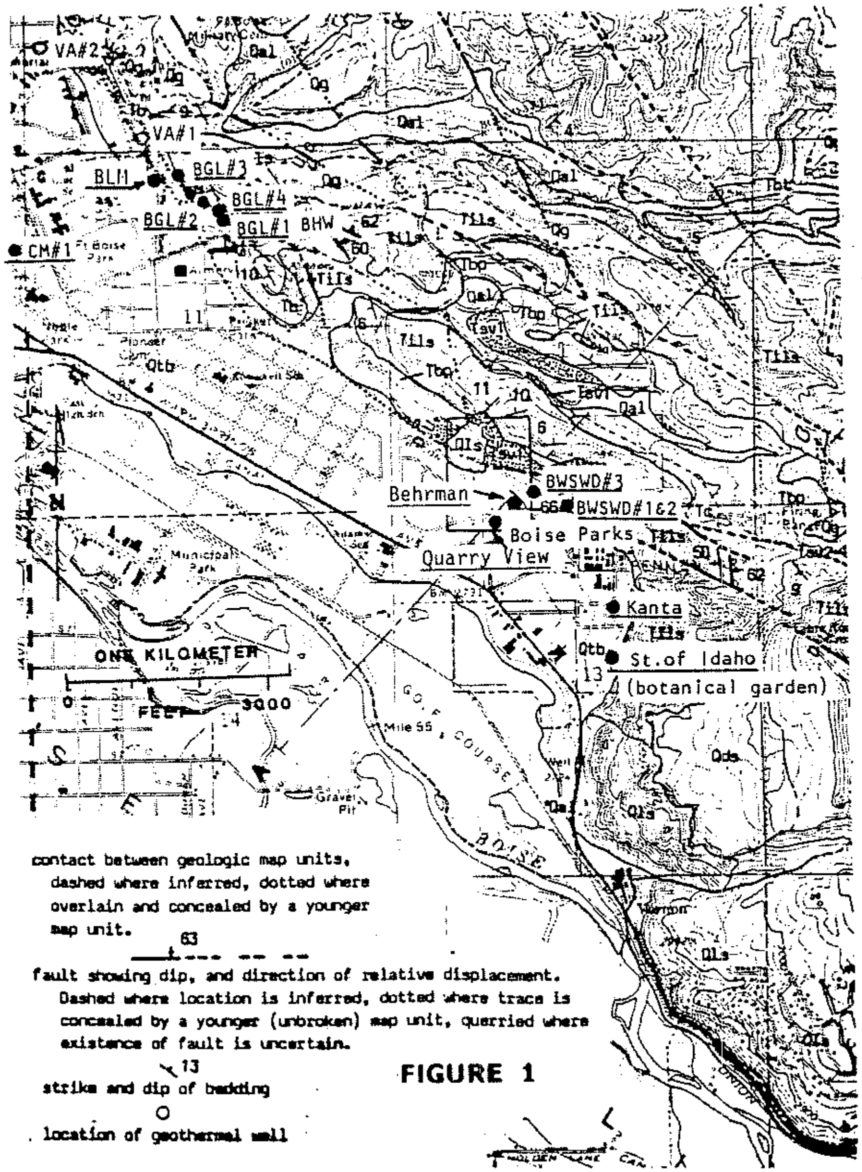


FIGURE 1

Qal	Alluvial deposits within floodplain of modern streams.
Qtb	Alluvial deposits of a late Quaternary terrace of the Boise River.
Qls	Landslide deposits ranging in age from late Quaternary to active in historic time.
Qds	Sandstone layers, folded and disturbed by shallow landsliding prior to the Warm Springs Mesa landslide.
Qg	Gravel deposits of stream terraces generally 16 m (50 ft) above modern streams
QTs	Sandstone, sand, gravel high in the stratigraphic section, but considered part of the lower Idaho Group.
Tils	Sandstone and minor silt layers of the lower Idaho Group.
Tb	Basalt within sediment of the lower Idaho Group.
Tbt	Basaltic tuff, brown in color with interbedded basalt flows.
Tc	Clay, maroon and light green in color with interbedded thin arkosic sand.
Tpb	Basalt, typically contains one or more flows with large plagioclase phenocrysts.
Tsv1	Rhyolite of "Castle Rock"
Tsv2	Rhyolite of Rocky Canyon.
Ki	Granitic rocks of the Idaho batholith.

Geothermal wells in northeast Boise. Geology after Burnham and Wood (1985). BWSWD #1 and 2 are producing wells. BWSWD #3, Kanta, and BLM were used as observation wells in this study. State of Idaho (botanical gardens) and Boise Parks are irrigation wells using geothermal water. BGL #2, 3 & 4 are producing wells. BGL #1, VA #1 & 2, & BHW are not producing at the present time. CM #1 is a reinjection well for the Capital Mall system.

waters are discharged into local canals or into the sewage system by the individual domestic users.

STATEMENT OF PURPOSE

Owing to the success of recent efforts to complete large capacity wells within the geothermal aquifer it was recognized that demands on the system may soon increase beyond the capacity of the natural system to recharge and/or repressurize the aquifer. Evidence of declining water levels and delays in the recovery peaks have further focused attention on a need for base-line data to determine the aquifer system's response to present and future withdrawals. Concern for prudent development of this important natural resource has led to efforts to monitor water production and water level pressures in available wells. This study, sponsored by the Idaho Department of Water Resources, is an effort to continue base-line data collection and its interpretation. Initiation of systematic data collection was begun in 1984 under a grant from the Idaho Water Resources Research Institute.

GEOLOGIC FRAMEWORK OF THE AQUIFER SYSTEM

The Boise Geothermal aquifer is a fractured media system. One favorable geologic setting for successful geothermal well completion is within the fracture network of a northwest-trending major fault zone which forms the boundary between the Snake River Plain and the Boise foothills (Fig. 1). That setting is here designated as a Type I setting. The fracture network exerts strong control on flow within the geothermal system. The fault zone apparently controlled natural discharge from the system in the late 19th Century manifested as seeps and springs of warm water (Lindgren, 1898; Wells, 1971). Most of these springs have ceased to flow over the years since production from the Boise Warm Springs wells began in 1892. In this setting along the base of the foothills, most large production wells derive their principal flow from fractures in rhyolite within the fault zone. Wells in this setting include the Boise Warm Springs Water District Wells, the Boise Geothermal Limited well field, and the Veteran's Administration wells (Figs. 2 and 3). Two recently drilled wells in the footwall of the zone did not yield enough water to justify completion. These are the Boise Warm Springs No. 3 and the Boise Geothermal Ltd. No. 1. In both cases, the wells were drilled through the normal fault zone and into the footwall without encountering sufficient fracture permeability. Most successful wells are located southwest of the surface trace of the fault zone in the hanging wall. Wells should be designed to penetrate a significant section of rhyolite in the fractured hanging wall.

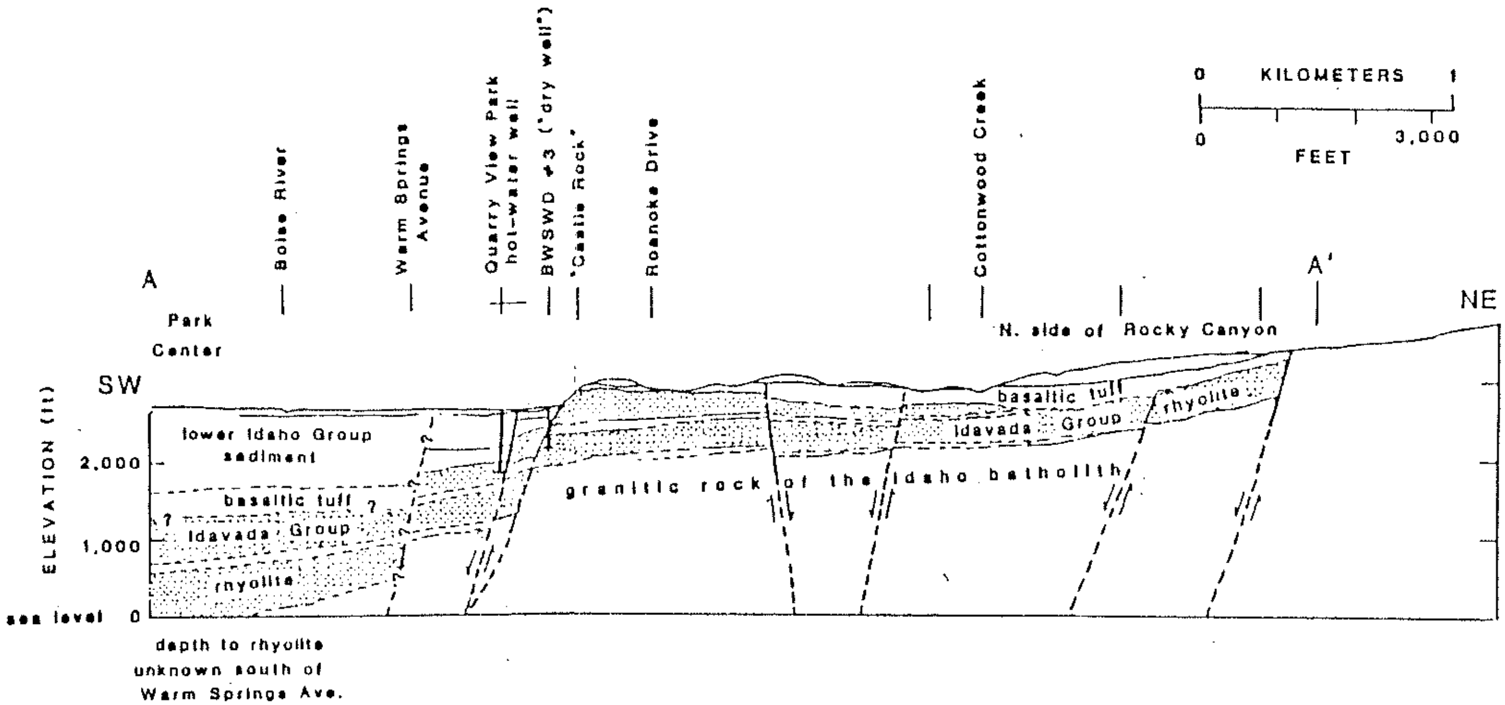


FIGURE 2. Geologic cross-section through the Boise Warm Springs area. (From Wood and Burnham, 1987, Fig. 2)

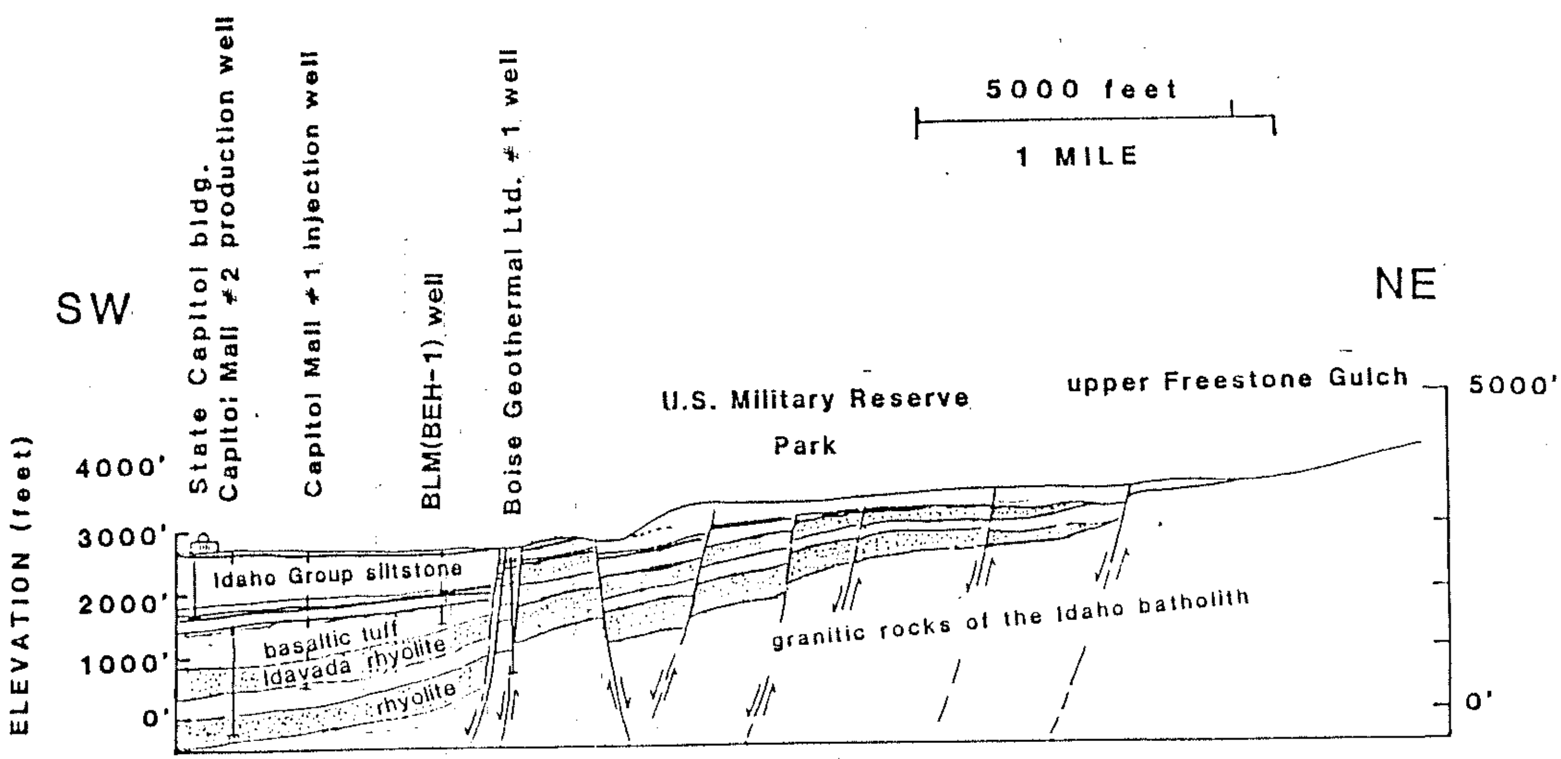


FIGURE 3. Geologic cross-section through the Boise Geothermal, Ltd. wells and the Capitol Mall wells. (Modified after Burnham and Wood, 1983, p.20)

A second geologic setting that is favorable for completion of large yield wells is within the tract of the Boise River Valley adjacent to and within a mile or so of the foothills. This zone is comprised of a series of hanging wall blocks which are generally downdropped toward the southwest and segment the aquifers of the Idavada Group. In the subsurface, the upper Idavada Group is composed of two rhyolite layers, each about 200-300 feet thick, separated by approximately 200 feet of sandstone and conglomerate. The total thickness of the rhyolite-sediment sequence is not known; however, the rhyolites appear to be widespread in the subsurface. The rhyolite layers are generally tabular, but thin to the northeast where the section laps upon granitic rocks of the Idaho batholith. Seismic reflection sections of the valley three miles east of the geothermal field show that the entire section of Miocene-Pliocene rocks dips about 6° to the west-southwest. Because of this structural tilting toward the center of the plain, and because of down faulting to the southwest, the rhyolite layers lie at greater depth to the west and southwest. The Capitol Mall wells are the most westerly exploration into the subsurface rhyolite section. The lateral and vertical extent and the structure of the rhyolite section is of interest if we are to eventually understand the nature of groundwater storage and recharge to the system. The upper 200 feet of the upper rhyolite is highly fractured. The fractured portion of the rhyolite is thought to be the principal aquifer in this Type II geologic setting. For instance, the majority of the

Capitol Mall No. 2 well is believed to come from this portion of the upper rhyolite. However, some production may be derived from the underlying 200 feet of sandstone and conglomerate, and the lower 300 feet drilled into the lower rhyolite.

Overlying and confining the rhyolite aquifer are 200 - 600 feet of volcaniclastic sediments and basaltic tuff. These are in turn overlain by 200 - 1,000 feet of claystone, siltstone, and minor sand units of the lower Idaho Group. Examination of rhyolite outcrops in the foothills, indicates that of all the rocks in the geologic section, the rhyolites are the only units with abundant, open, unhealed fractures. The origin and nature of the fractures has not been determined. A study would be useful to determine if the fractures are widespread in the upper portions of the rhyolite layers such as original cooling fractures, or if they are confined to fracture zones such as would be expected of tectonic fractures. Wells in this geologic setting are not associated with faults identified by surface geologic mapping or by subsurface correlation of well logs, they include the Capitol Mall, Kanta, Quarry View Park, and the BLM (BEH) wells. Of these, only the BLM well was unsuccessful. It experienced a drawdown of 169 feet with a pumpage of 120 gpm for 30 hours (Nelson and others, 1980). An accurate log of the well is not available, but it appears that the well did not penetrate the rhyolite. If it did, it is likely that only the uppermost part of the rhyolite was penetrated; perhaps, that may be the reason for the poor production.

A third but less common geologic setting conducive to production of geothermal water in the Boise area is from the granite of the Idaho batholith (Type II). Success of the Harris well drilled in 1980 indicates that open fractures may be encountered in faulted granitic rocks, as well as in the rhyolite.

The Harris well is the most easterly hot water well of the system. The well produces 180°F water, the hottest in the system, from fractures in sandstone and granite along the foothills fault zone near the site of a former warm springs. No rhyolite was encountered in that 890-ft well (interpretation by S. Wood, based on examination of cuttings).

Although portions of the rhyolite sequence are clearly important aquifers, inclined fault zones in the granitic, sedimentary and volcanic rocks are also a part of the ground-water storage and circulation system. Fault zones having significant vertical permeability, serve as conduits for deeply circulating groundwater, whereas the tabular rhyolite layers and fracture zones appear to provide the system with lateral hydraulic conductivity, and interconnections within the system.

Orientation of faults within the fracture system varies from N60°W to north-south. Faults are of different ages, some faults offset older units only and do not cut younger units. Measured attitudes of surface outcrops of fault planes within the foothills fault zone strike N70°W to N45°W. Dips range from 55° to 70° and are predominantly to the southwest. A smaller population of minor faults are antithetic to the main fault. They are downthrown and dip to the northeast, and form a number of small graben and horsts within the main fault zone. Many of the faults mapped in the foothills probably project under the Valley, but their exact position and degree of interconnection is not well known.

The rhyolite and sedimentary section is faulted against granite in several places in the foothills by these northwest-southeast trending fractures. Faults within the granitic rocks of the batholith have not been documented by detailed mapping. However, numerous warm springs elsewhere in the batholith are evidence that deep groundwater circulation occurs in fracture zones in the granitic rocks. As yet we know little of their geometry or if those springs are somehow related to the Boise Geothermal System.

Table I lists the wells which penetrate the geothermal aquifer in the Boise Vicinity. Where the information is available, the total depth, formation, and geologic setting in which the well is completed or terminated are indicated.

TABLE I

GEOTHERMAL WELLS IN THE BOISE VICINITY

WELL NAME	TOTAL DEPTH (FEET)	FORMATION IN WHICH TERMINATED OR COMPLETED	GEOLOGIC SETTING
Behrman	?	?	?
BGL No. 1	2,000	IDAVADA	Type I (Fault Zone)
BGL No. 2	880	IDAVADA	Type I (Fault Zone)
BGL No. 3	1,897	IDAVADA	Type I (Fault Zone)
BGL No. 4	1,103	IDAVADA	Type I (Fault Zone)
BHW	1,281	?	Type I (Fault Zone)
BLM (BEH)	1,224	Basaltic Tuff	Type I (?) (Proximal to Fault Zone?)
BWSWD No. 1 & 2	410	IDAVADA	Type I (Fault Zone)
BWSWD No. 3	595	IDAVADA	Type I (Fault Zone)
CM No. 1	2,152	IDAVADA	Type II (Fractured Rhyolite
CM No. 2	3,030	IDAVADA	Type II (Fractured Rhyolite
Edwards Greenhouse	1,195	?	Type I (Fault Zone?)
Harris	690 (approx)	Idaho Batholith & Idaho Group?	Type III (Fractured Granite?
Kanta	1,105	IDAVADA	Type II (Fractured Rhyolite
Koch	1,143	?	?
Milstead	?	?	?
Quarry View	865	IDAVADA	Type II (Fractured Rhyolite
State of Idaho (Botanical Garden)	875	Idaho Group of IDAVADA	Type II (?) Fractured Rhyolite
State of Idaho	487	IDAVADA	Type II (?)
Veterans Administration			
(test)	1,847	IDAVADA	Type I (Fault Zone)
(production)	1,666	IDAVADA	Type I (Fault Zone)
(Injection)	2,312	IDAVADA	Type I (Fault Zone)

GROUNDWATER AQUIFERS OVERLYING THE GEOTHERMAL SYSTEM

DEEP GROUNDWATER SYSTEM

The "deep groundwater system" as defined by Dion (1972), Nace and others (1957), Ralston and Chapman (1970), and Burnham (1979), is from a depth of 200 feet to 800 feet. The stratigraphic section is mostly silty claystone, and siltstone of low permeability, but which contains several important beds of sand and sandy gravel. As the stratigraphy of the valley is presently understood, most of these sediments are within the lower part of the Idaho Group (Wood and Burnham, 1983) and not part of the Glenns Ferry Formation as originally reported by Dion (1972). The Glenns Ferry Formation should be regarded as the upper Idaho Group and is recognized by Wood and Burnham (1983) as a sand facies capping the western Boise Foothills. Geologic formations within the Idaho Group beneath the Boise River Valley have not been studied in sufficient detail to use formation subdivisions at this point. Burnham (1979) describes the deep system as overlain by a clayey sequence of low permeability which restricts water movement rather than forming an absolute separation. Sand aquifers within this clayey sequence are the source of the major drinking water supply for the City of Boise and other municipalities in the area. The "deep groundwater system" is thought to recharge near the Boise River, east of Boise, and to discharge to deep wells and to the Snake River toward the south and southwest (Ralston and Chapman, 1970; Young, 1979). The recharge characteristics have never been examined in detail. The extent to which the deep part of this system may interconnect along faults with the geothermal system is not known. Most wells deeper than 500 feet are warm, 60-80^{o}F, but that may be due entirely to the local geothermal gradient, and not to intermixing with water from the geothermal system. Mixed water would clearly be indicated by anomalous fluoride-ion content (F^{-}>1.0mg/liter) because waters from the principal geothermal wells exceed 15 mg/liter fluoride. Fluoride content of deep wells in the valley needs to be examined, and that is a problem for additional research.

SHALLOW GROUNDWATER SYSTEM

The shallow groundwater system as defined by Dion (1972) and discussed by Burnham (1979) is that continuous body of groundwater that lies within the upper 100 to 300 feet of sediments of the Idaho Group, fluvial sediments of the Boise River, and basaltic lavas of the Snake River Group. This groundwater system is under essentially free water table conditions and responds seasonally to recharge by precipitation and mostly to imported irrigation water. The water-level map for October, 1970, constructed by Dion (1972) shows the New York Canal as a major source of recharge. The water table slopes to the west-northwest about 20 ft per mile. More recently, septic tank and sewage treatment effluent are a component of recharge. Water from the shallow groundwater system discharges in west Boise to individual irrigation and domestic wells and by drainage ditches and springs to the Boise River. The water table of this shallow system rose rapidly in the 1912-1915 interval as new irrigation waters were supplied by canals of the Boise Project. The water table is as deep as 300 feet beneath the upper terraces in the southeast Boise area, but is very near the surface in west Boise. Local perched tables occur within this upper section and are discussed most recently by Burnham (1979).

HYDROLOGIC SETTING AND DATA GATHERING

BOISE WARM SPRINGS WATER DISTRICT (BWSWD) WELLS 1 and 2
(Elevation of Pumphouse Floor - 2,764.9 feet)

Boise Warm Springs Water District pumping well, (BWSWD) Wells Nos. 1 and 2, and an observation well, BWSWD No. 3, are shown in Figure 1. Wells 1 and 2 are 30 feet apart and are both drilled into a fracture zone which strikes approximately N70°W, dips steeply southwest and is downthrown to the southwest. Well No. 2 is the principal pumping well for the district. Usually No. 1 is pumped only when No. 2 cannot supply the demand on the system. Figures 4 and 5 show drawdown versus time in wells 2 and 1, respectively, plotted with combined production in thousands of gallons per minute (kgpm) from the system. The periods for the plots are 6/11/85 to 6/11/86. The data for the plots were provided by the Boise Warm Springs Water District Engineer, Mr. Robert Griffiths. Water-level measurements are made using an airline in each well. Flow-rate measurements are calculated from a Sparling cumulative flow meter installed in a common pipe served by both wells. Water levels and the Sparling meter are read once per day, usually in the morning between 8 a.m. and 10 a.m., or occasionally in late morning or afternoon. The readings reflect the water level only at the time of reading and are somewhat skewed toward levels at a time of day when domestic demand is high. However, during the peak heating season, even random readings commonly reflect a maximum drawdown.

FIGURE 4 BOISE WARM SPRINGS WELL NO. 2
(WEST WELL. EL. 2,764.9 ft)

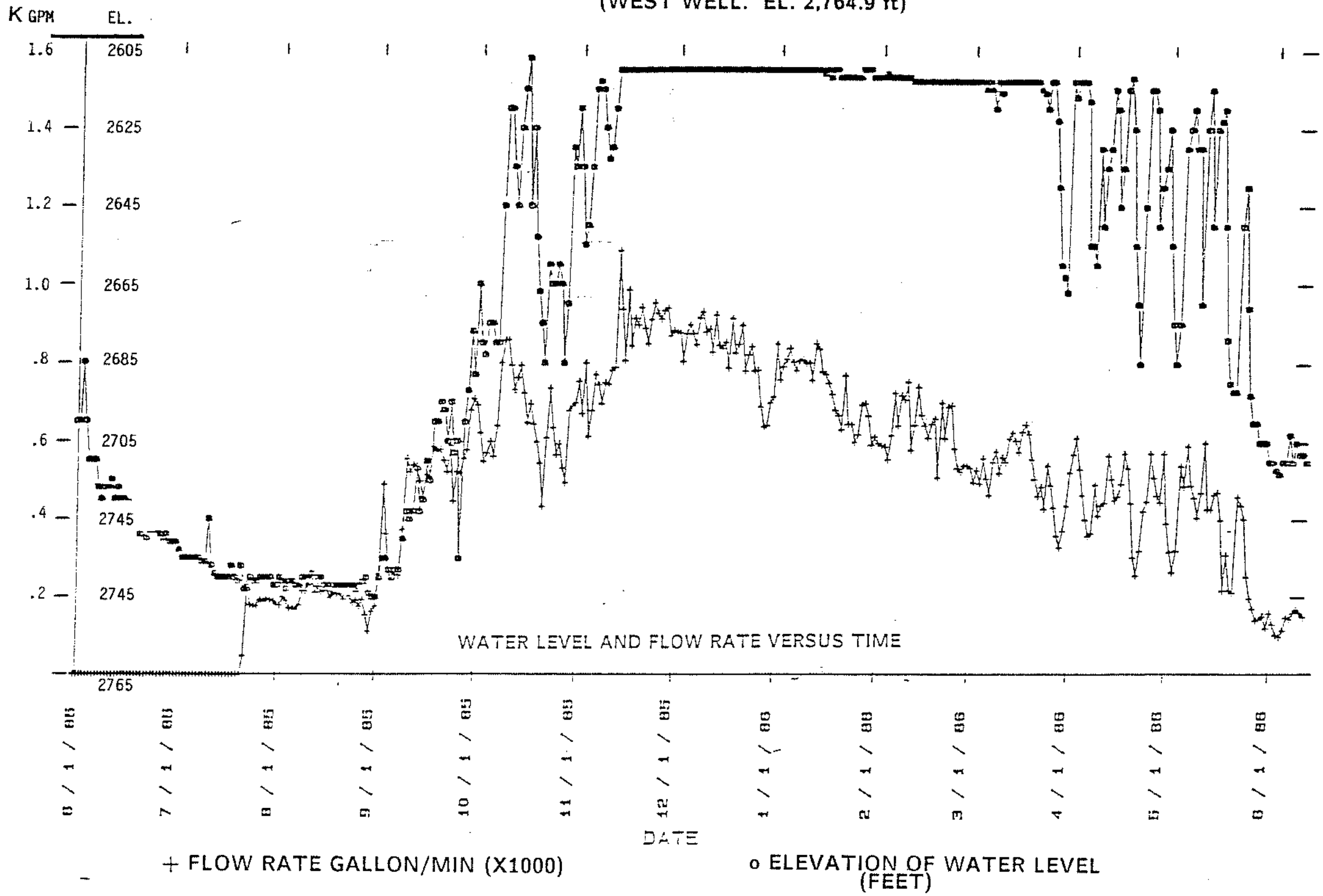
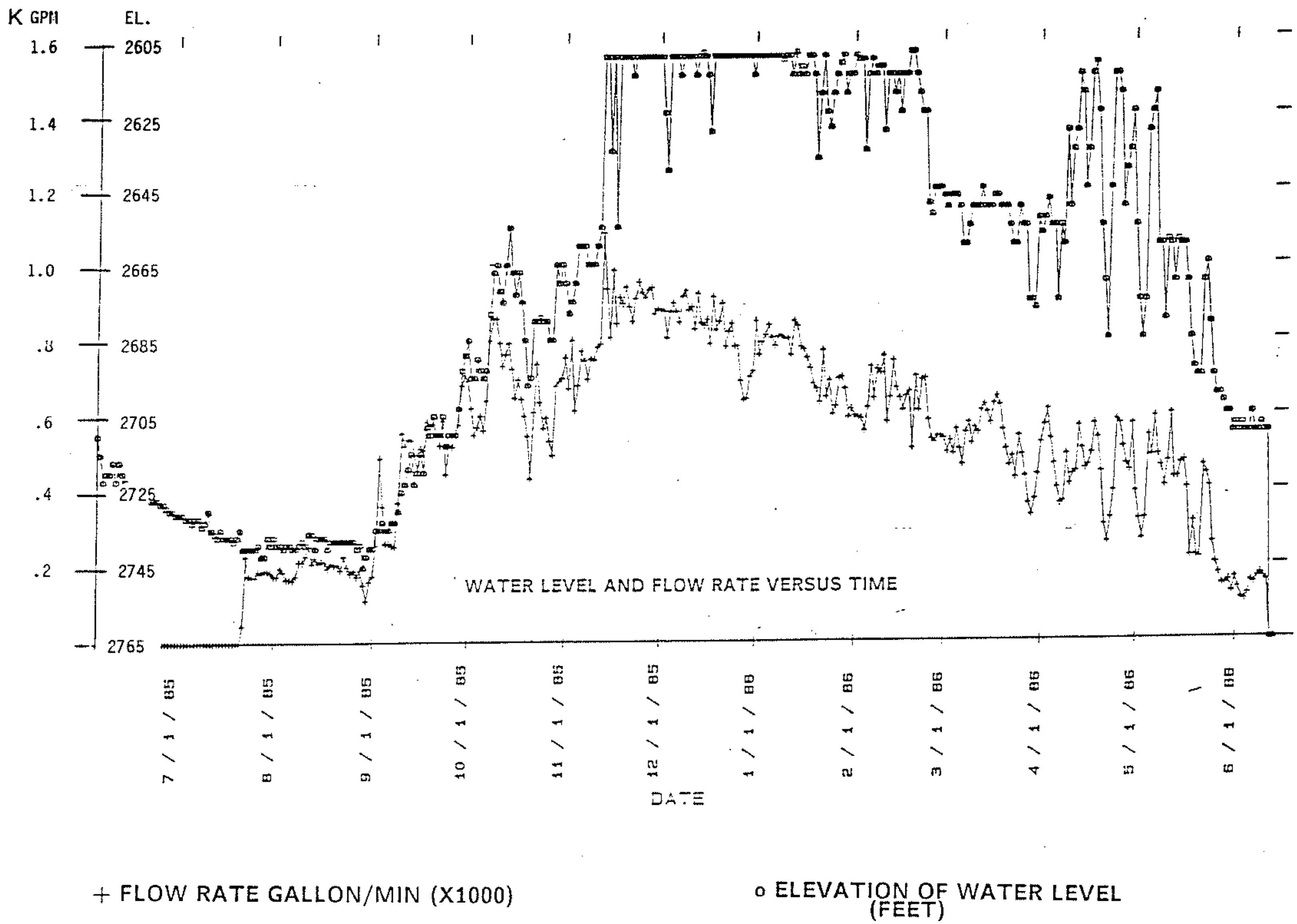


FIGURE 5 BOISE WARM SPRINGS WELL NO. 1
(EAST WELL, EL. 2764.9 ft)



The Sparling meter readings are averaged over the period since the previous reading was taken (approximately 24 hours) to determine the pumpage in gallons per minute; thus, they reflect an average flow rate for that period. Unfortunately, a gap exists in the flow-rate data. During the period 6/11/85 to 7/22/85 the Sparling meter was under repair. Fortunately that was a period of low demand as is indicated by the continued recovery of the water levels. The measured pumpage from BWSWD Wells 1 and 2 during the period 7/22/85 to 6/11/86 was 254,340,000 gallons. To estimate the withdrawal during the period of missing data 6/11/85 to 7/22/85, the past records were used. The only years in which production for that entire interval was recorded were 1979 and 1980 when 14.6 million and 13.7 million gallons respectively were pumped. Using an average of these two years of 14.3 million an estimate of the total pumpage from Wells 1 and 2 of approximately 268.6 million gallons is obtained. This yields an average pumping rate for the year of 510 gpm. These estimates are probably within 2-3% of the actual withdrawals by Wells 1 and 2.

The elevation of the BWSWD pumphouse floor is 2764.9 feet. Water levels within wells 1 and 2, the pumping wells, did not recover sufficiently to flow at the surface again in 1985. The maximum recovery within those wells was to within 15 feet of the pumphouse floor and occurred during late August 1985. It is worthy of note, that surface flow formerly common to the wells has not occurred since the summer of 1983. In the fall of 1985 drawdown in the principal pumping well, No. 2, occurred rather abruptly beginning in early September (Fig. 4) and by October 15, 1985 declined temporarily below the level of the pump bowls to a depth of 158 feet (Pump bowls are approximately 150 feet below surface. A 10-foot long tail pipe is installed below the pump bowls). This abrupt decline reflects the cool to cold temperatures in the fall of 1985. By November 12, 1985, the water level in well No. 2 had again declined to the level of the pump bowls. This time the water level remained at or near its maximum drawdown until March 25, 1986, a period of 133 days. To meet demand, well No. 1 began pumping on November 12. The water level in well No. 1 on November 12th was 110 feet below the pumphouse floor at an elevation of 2654.9. By November 13th, both wells were pumping from near their maximum drawdown levels of 155 feet at an elevation of 2610 feet (Figs. 4, 5). The pumping level in No. 1 remained at or near its maximum until mid-February. During the 133 day period in which drawdown was near its maximum, pumpage ranged from a high daily average of 1085 gpm in mid-November to a low daily average of 430 gpm in late March. The average pumpage rate from both wells for the 133 day period was 725 gpm.

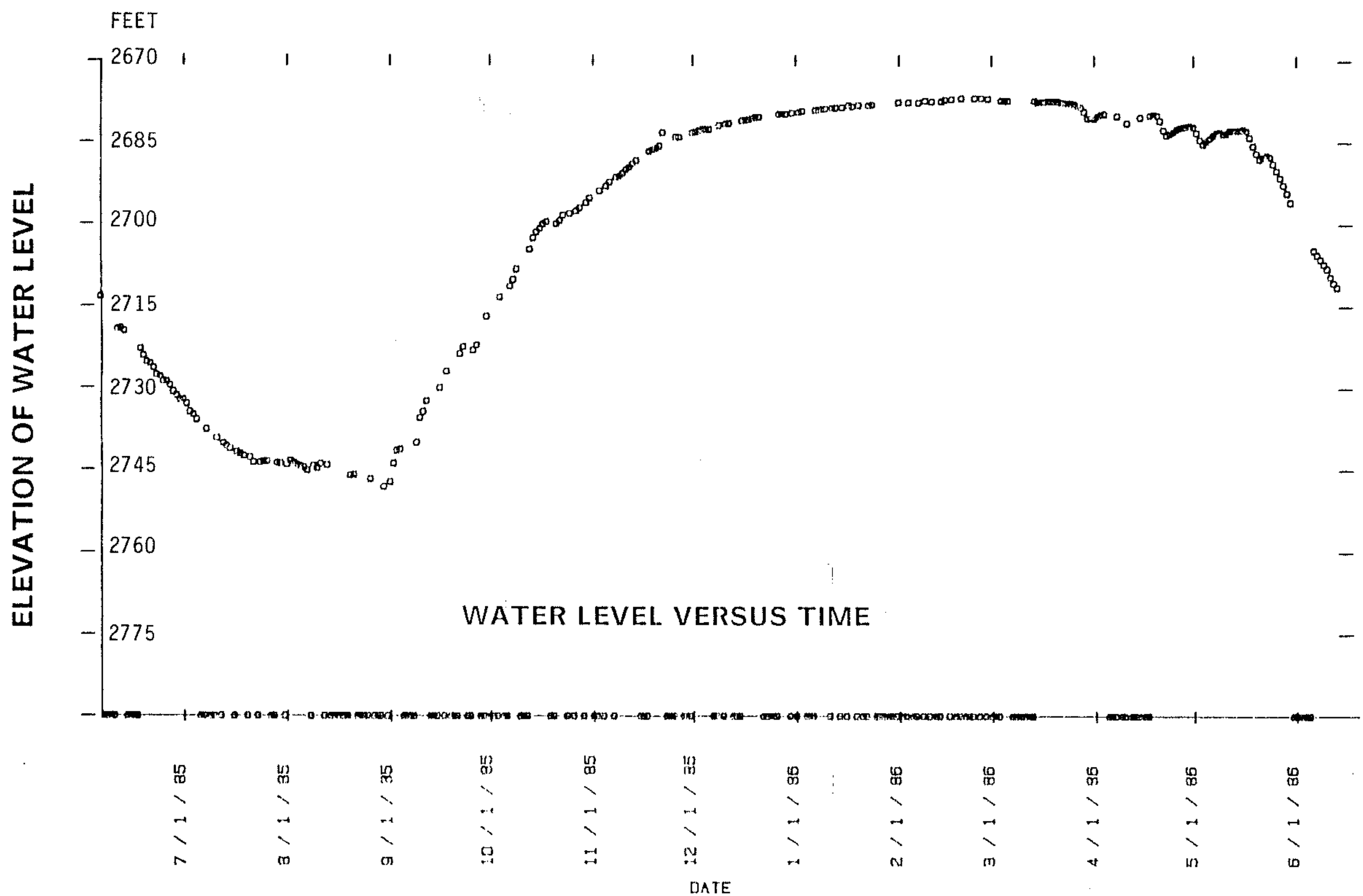
BOISE WARM SPRINGS WATER DISTRICT WELL NO. 3
BWSWD OBSERVATION WELL NO. 3
(Ground Elevation 2787.54 feet)
(Top of Casing 2789.5 feet)

Maximum recovery in BWSWD well No. 3 occurred on August 30, 1985, at an elevation of 2747.5, 42.0 feet below the casing collar. In response to withdrawals by BWSWD wells 1 and 2, the water levels in well No. 3 began to decline again by September 1, 1985. The decline in the water level in well No. 3 in late 1985 and early 1986, was well ahead of the same period in 1984-85. The well finally reached its lowest level on February 25, 1986, at an elevation of 2676.5 ft., 113 feet below the collar and at a maximum drawdown of 71.2 feet (Fig. 6).

During this study water levels were determined by frequent tape measurements and by use of a Stevens "F" type recorder. Unfortunately, even though a cyclone fence was installed around the well, and the recorder was covered with a steel box, the well was vandalized resulting in some lost records.

Comparison of Figure 6 to the drawdown curve of the principal pumping well, BWSWD No. 2 (Fig. 4) indicates that the water level in well No. 3 moves in rather direct response to water level changes in the principal pumping well. As is to be expected, the fluctuations of the water level in well No.3 are more subdued because well No.3 is approximately 645 feet northwest of the pumping wells. The three BWSWD wells seem to be well-connected hydraulically. The interconnection is indicated by the similarity of their drawdown curves, and by what appear to

FIGURE 6 - BOISE WARM SPRINGS WELL NO. 3
(EL. of Casing Collar - 2,789.5 ft)



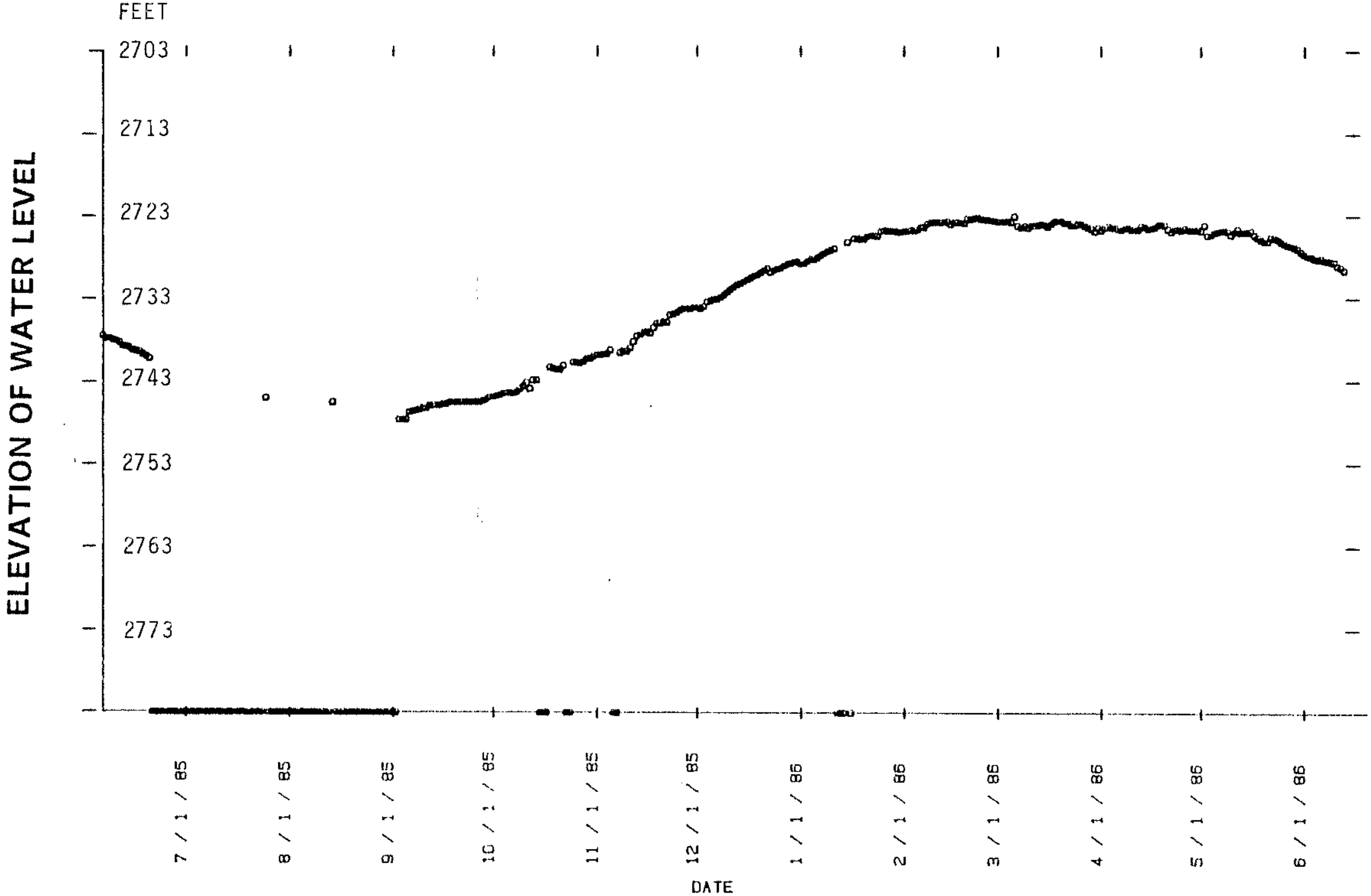
be rapid responses recorded in well No. 3 to changes in pumpage in well No. 2. The good hydraulic connection between the wells reinforces the interpretation that No. 3 is drilled into the same fracture zone as the pumping wells. However, drilling for No. 3 apparently continued through the fracture zone into the footwall and the hole was terminated at 600 feet in rhyolite. Pump tests on the well at the time of drilling did not yield sufficient water to justify completing the hole for production, but it is an important observation well for the geothermal system.

KANTA WELL
(Observation)
(Ground Elevation 2,782 feet, approximate)

The Kanta well is approximately 1,675 feet south 10° east of the BWSWD pumping wells (Fig. 1). The well was drilled from a surface elevation of approximately 2,782 feet to a total depth of 1,015 feet. The elevation of the casing collar is about 2,783 feet, and depth to water-level measurements are given with respect to the collar. Until October 11, 1985, a Steven's "A" type recorder was used to monitor water levels within the well. However, owing to frequent incidences of vandalism, damage to the recorder resulting in a loss of record, and a desire for greater sensitivity, an "F" type recorder was installed.

The water-level in the Kanta well reached its peak recovery on September 3, 1985, at an elevation of approximately 2,747.5 or 35.38 feet below the casing collar, Figure 7. The 1985 recovery

FIGURE 7 - KANTA WELL
WATER LEVEL VERSUS TIME
(Approximate EL of Casing Collar - 2,783 ft)



peak was approximately 2 feet below the previous year, which peaked at an elevation of 2,749.5 on September 20, 1984. The water-level reached its lowest level on February 21, 1986 at an elevation of approximately 2,723.4 feet or 59.6 feet below the collar. This compares with its lowest level on February 27, 1985 at 2,728.6 feet. This year's maximum decline was about 5 feet below that of last year.

Best estimate is that the Kanta Well is not completed within the fractures that include BWSWD wells 1, 2, and 3, but is drilled into a down-thrown fault block that lies southwest of that fault zone. However, the Kanta well, as can be seen in Figure 7, responds very well to withdrawals by the Warm Springs pumping wells. Comparison of Figures 4, 5, 6, and 7, again shows a similarity of curves. Some of the abrupt water-level fluctuations in the Warm Springs pumping wells seem to be recognizable in the curves of both well No. 3 and the Kanta well. The fluctuations however, are attenuated in the observation wells; more so in the Kanta than in well No. 3, because of its greater distance from the pumping wells.

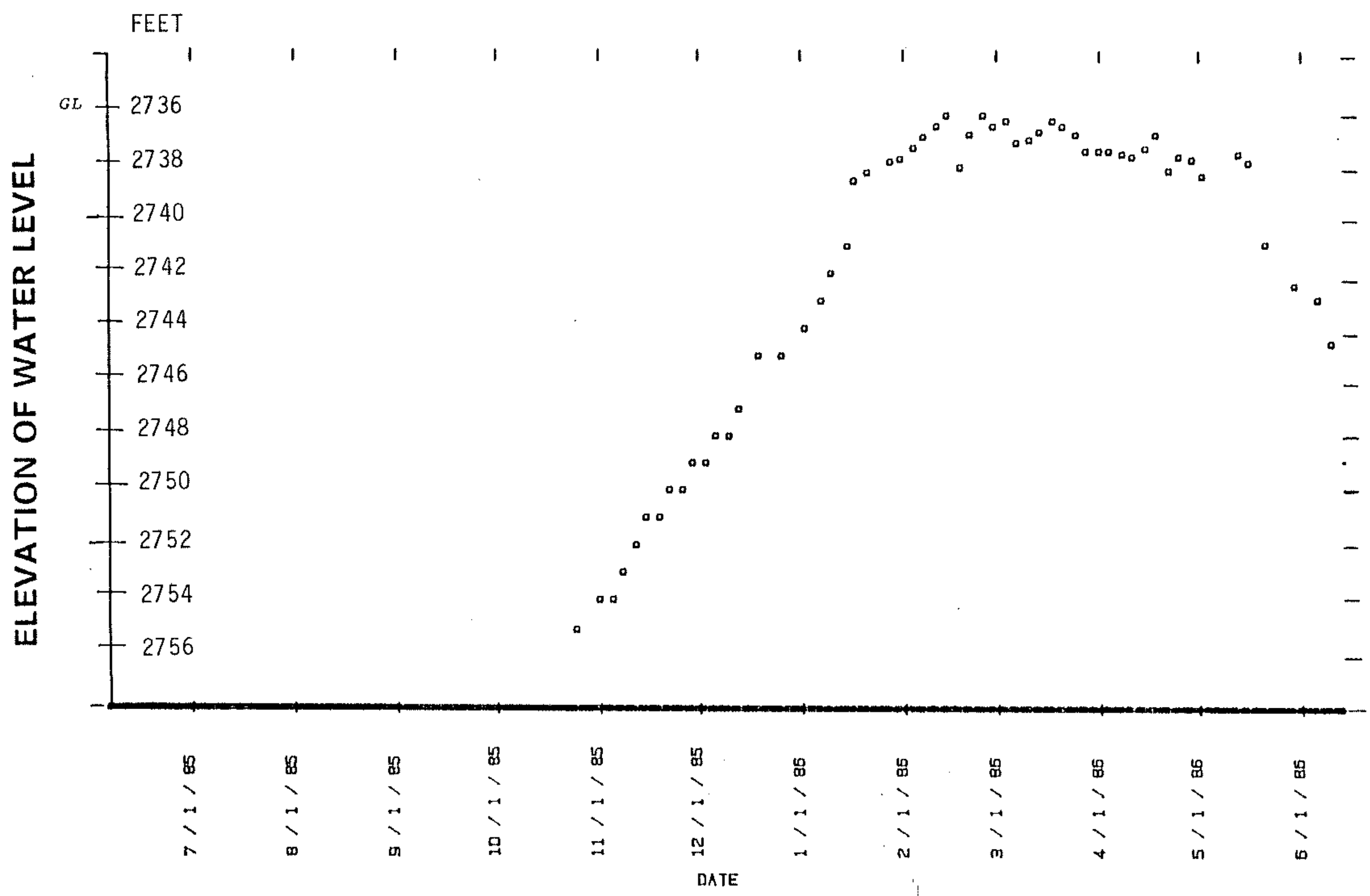
BEHRMAN WELL
(Observation)
(Elevation 2,735.7 feet)

The Behrman well is approximately 720 feet west of the BWSWD pump house in a marshy area at an elevation of 2,735.7 feet. The well is old and is drilled to an unknown depth. Geothermal water commonly flows at the surface under pressure in contrast to the deeper water-levels typical of BWSWD wells 1, 2, and 3 in recent years. To attempt to understand the hydraulic relationship between the Warm Springs wells and the Behrman well, a twice-weekly monitoring of the Behrman well was begun on October 25, 1985.

The Behrman well is not secured by a fenced enclosure and pressure gauges are subject to vandalism. Therefore, the following monitoring procedure was followed to obtain data from the well. When the well was flowing, a pressure gauge was attached to the well and the pressure was allowed to build for 10 minutes, at which time the pressure is recorded. Experience indicates that the well pressure will be essentially stabilized within 10 minutes during periods of positive pressure at the well.

Data presented in Figure 8, indicate that the Behrman well maintained a positive pressure at the well head until January 15, 1986. The head elevation at that time was about 2,739 feet, 3 feet above the ground elevation. The lowest water level measured in the Behrman well during the monitoring period was on February 25, 1986 at approximately 3 feet below the well head, or at the

FIGURE 8
BEHRMAN WELL
(Ground Elevation - 2,735.7 ft)



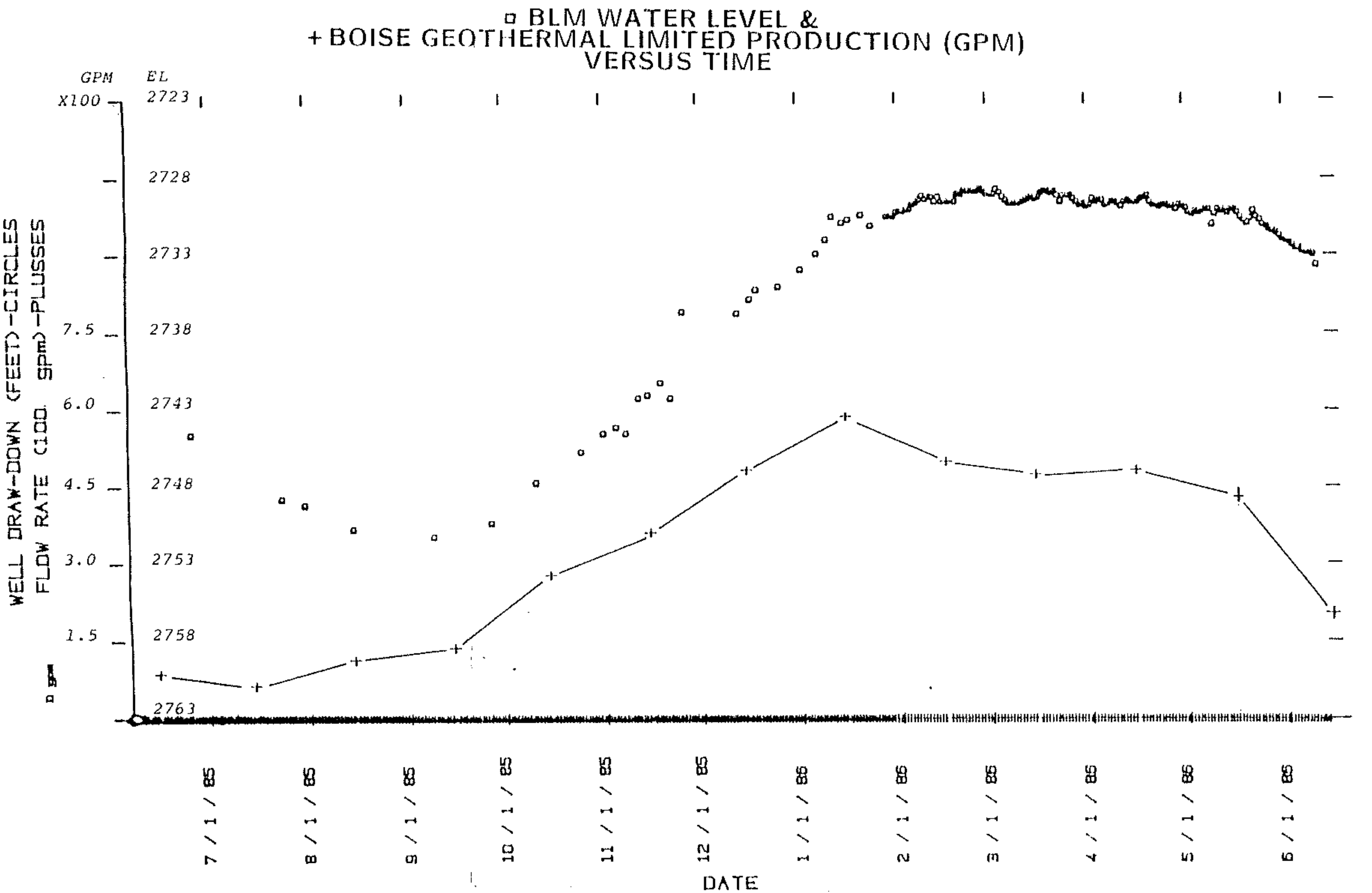
ground elevation of 2,736 feet.

BLM (BEH) WELL
(Observation)
(Ground Elevation 2,742.7 feet)

Monitoring of the BLM Observation well began on June 4, 1985, using a pressure gauge installed in a port in the casing at an elevation of approximately 2,739.7 or 2.5 feet below the ground reference elevation. In early December 1985, the water-level in the observation well declined below the port. From that time until early February, measurements were made using an electric tape entering the well through that same port. In late January 1986, the U.S.G.S. Water Resources Division, extended the casing on the well to an elevation of 2,745.5 feet, erected a shed over the well, and installed a Stevens "A" type recorder. On February 4, 1986, the well was equipped with a Campbell Data-Logger and Druch pressure transducer. The data-logger and "A" recorder readings were checked twice weekly.

The maximum recovery observed in 1985 was recorded on September 4, at 4.9 psi or 11.3 ft. above the gauge at an elevation of 2,751 ft., Figure 9. The lowest water level recorded during this study period occurred on February 26, 1986, at 14.7 feet below land surface, at an elevation of 2,728 feet. On September 10, the well reached its maximum recovery for the 1986 recovery period of 2,747.15. This is approximately 4.5 feet above the ground elevation of the well, but 4 feet short of its

FIGURE 9



□ BLM WATER LEVEL &
+ BOISE GEOTHERMAL LIMITED PRODUCTION (GPM)
VERSUS TIME
GPM X100
EL
2723
2728
2733
2738
2743
2748
2753
2758
2763
7.5
6.0
4.5
3.0
1.5
WELL DRAW-DOWN (FEET)-CIRCLES
FLOW RATE (100. gpm)-PLUSSES
7 / 1 / 85
8 / 1 / 85
9 / 1 / 85
10 / 1 / 85
11 / 1 / 85
12 / 1 / 85
1 / 1 / 86
2 / 1 / 86
3 / 1 / 86
4 / 1 / 86
5 / 1 / 86
6 / 1 / 86
DATE

1985 recovery, Figure 10.

CITY OF BOISE
QUARRY VIEW PARK WELL
(Observation)
(Ground Elevation 2,728 feet, approximate)

Water-level measurements in the Quarry View Park well were begun on March 21, 1986. The water level within that well appears to have been essentially stabilized around 35 feet below the pumphouse floor until mid-May, Figure 11. Since that time the well has been pumped to supply irrigation water for the city park. The well was monitored briefly in March of 1985, before the well was cleaned out and placed into production, and at that time, the water level fluctuated between 32 and 38 feet below the land surface.

Judging from data currently available, the larger fluctuations recorded this year seem to be primarily related to pumping in the well and not to drawdowns imposed by other pumping wells. Drawdowns within the well due to its own pumping may be expected to be quite large, depending on the duration of pumpage. Pump tests were performed on the well in February of 1984, and indicate that rather rapid drawdowns occurred during an 8-hour test. Those pump test data are discussed in the analysis section of this report.

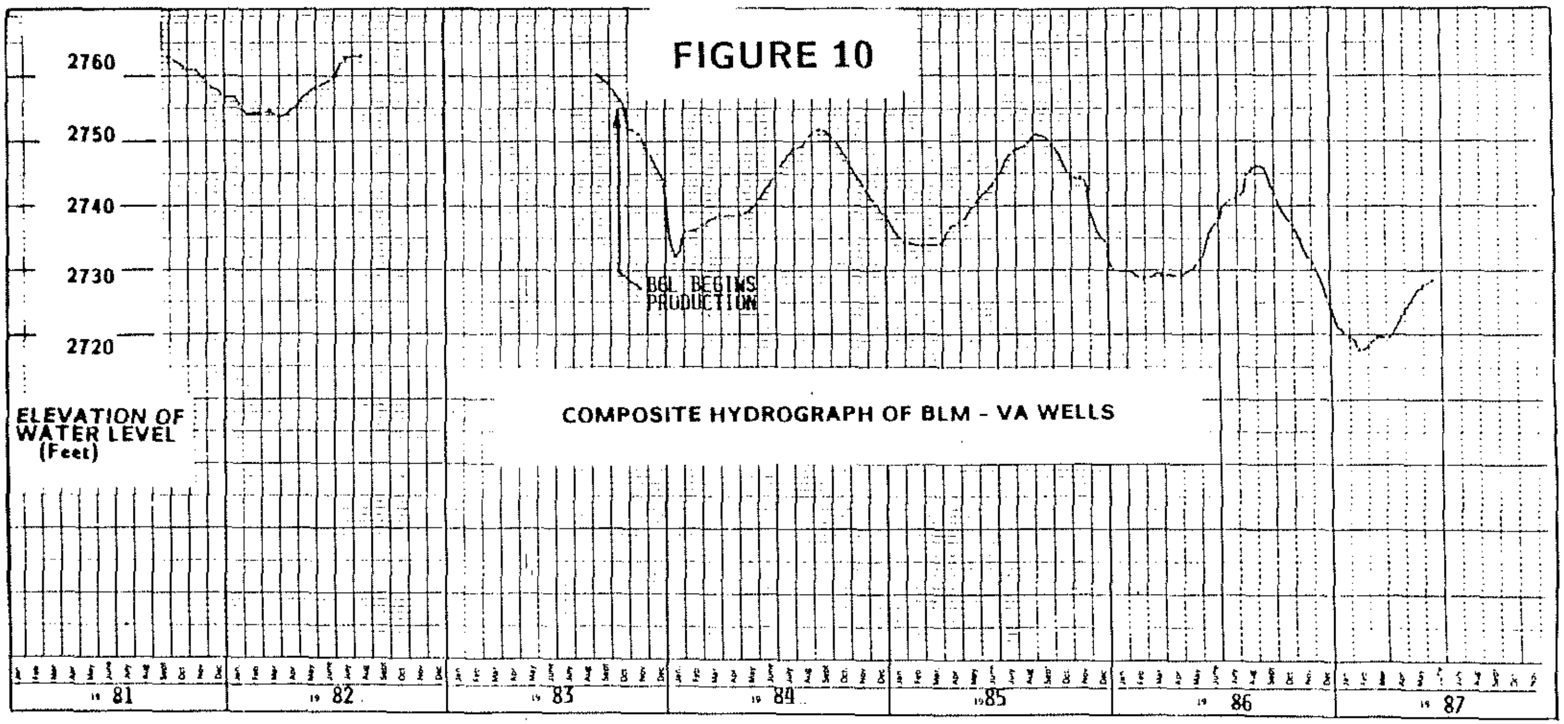


FIGURE 10
2760
2750
2740
2730
2720
ELEVATION OF WATER LEVEL (Feet)
BGL BEGINS PRODUCTION
COMPOSITE HYDROGRAPH OF BLM - VA WELLS
81
82
83
84
85
86
87

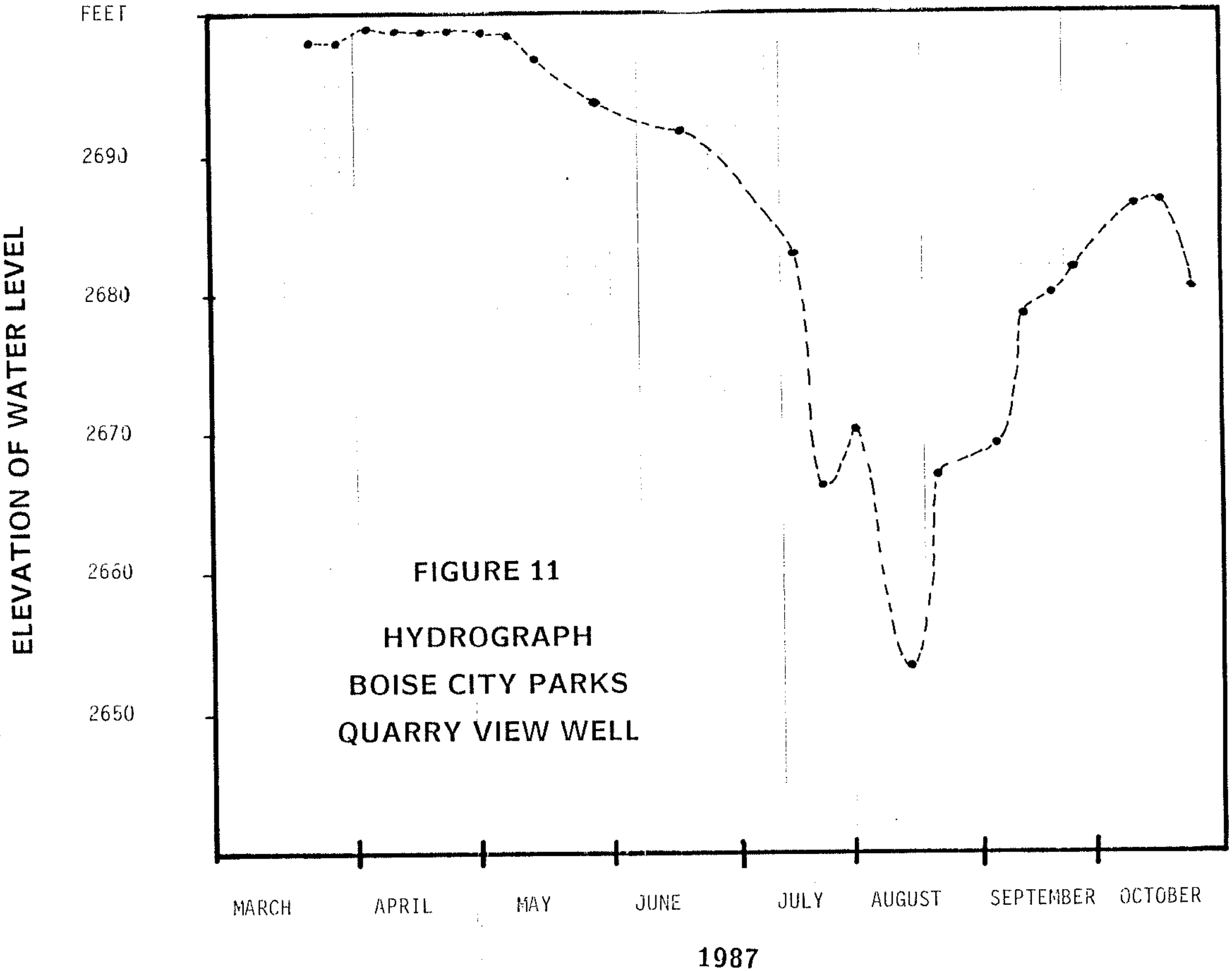


FIGURE 11

HYDROGRAPH
BOISE CITY PARKS
QUARRY VIEW WELL

ANALYSIS

BOISE GEOTHERMAL LIMITED - BLM (BEH) HYDROGRAPH FLUCTUATION

As indicated in Figure 1, the BLM Observation well is in close proximity to the Boise Geothermal Limited (BGL) wells. It is interesting to note that during the heating season, on week days between the hours of 6 am and 6 pm hydrographs from the BLM well show rapid fluctuations. These fluctuations are interpreted as responses of the BLM well to pumpage by the BGL well system. Although the individual fluctuations are small, less than one foot, they are abrupt and reflect good hydraulic interconnection between the observation well and BGL pumpage wells. Unfortunately, water level measurements from the BGL system are not available so no attempt is made to correlate the fluctuations within the BLM well to drawdowns within the BGL wells as they pump. Obtaining those water level data is the next logical and critical step in studying this portion of the system.

Composite monthly totals of withdrawals from the Boise Geothermal Limited wells are presented in Table I. Figure 9 shows those monthly withdrawals expressed in average gallons per minute for each month plotted with a hydrograph of the data from the BLM well for the period 6/1/85 to 6/1/86.

The hydrograph shows the recovery of the water level in the BLM well during the period June 1 until September 4, 1985, when the well reached its maximum recovery level at an elevation of

2,751 feet. After September 5, the water level in the BLM well declined until February 26, 1986 when it reached its lowest level at an elevation of 2,728 feet. The close parallelism of the drawdown and production curves in Figure 9, in addition to the already noted prompt response of the BLM well to short time pumpage in the BGL wells indicates that the BGL well activity is principally responsible for the current water level fluctuations in this portion of the system and that no significant impermeable barriers lie between the BGL production wells and the BLM well.

TABLE II

MONTHLY SUMMARY OF GEOTHERMAL WATER WITHDRAWN BY BOISE GEOTHERMAL LIMITED WELLS*

PERIOD	TOTAL GALLONS PUMPED	YEARLY TOTAL (GALLONS)	YEARLY TOTAL (ACRE FEET)
October 1983	12,894,700		
November 1983	17,565,300		
December 1983	23,306,400		
January 1984	32,091,300		
February 1984	16,187,700		
March 1984	12,646,000		
April 1984	13,226,810		
May 1984	13,543,570		
June 1984	7,367,020		
July 1984	5,527,710		
August 1984	5,898,120		
September 1984	6,441,750	166,696,380	511.6
October 1984	9,971,100		
November 1984	15,237,220		
December 1984	15,136,520		
January 1985	18,188,710		
February 1985	17,822,100		
March 1985	12,071,100		
April 1985	8,308,450		
May 1985	7,384,000		
June 1985	3,752,000		
July 1985	2,746,000		
August 1985	4,897,000		
September 1985	5,869,000	121,383,200	372.5
October 1985	11,894,000		
November 1985	15,477,000		
December 1985	20,739,000		
January 1986	25,295,000		
February 1986	21,459,000		
March 1986	20,318,000		
April 1986	20,752,000		
May 1986	18,751,000		
June 1986	8,730,600		
July 1986	1,279,350		
August 1986	861,380		
September 1986	11,267,070		
		176,823,400	542.7

*Data provided by the City of Boise Geothermal System, Department of Public Works.

Figure 10 is a hydrograph of the BLM well for the period September 1981 to October 1986. The data for the period 9/83 through 8/84 is a composite record of water-levels in the BLM well and the Veterans Administration well. All water-level data prior to June of 1985, are replotted from graphs in the Phase II Proposal - Boise City Geothermal Project (p.29). It is again evident that a close correlation between the plotted water-levels and the BGL pumpage exists, and extends back as far as October 1983 when extractions by the BGL wells began. The period October 1983 through September 1984, during which BGL extractions were large, 166.7 million gallons, was also the year of the greatest drawdown in the BLM well. Looking at the overall trend, it is also evident that the water-level in the BLM well shows a cumulative decline in maximum recovery levels from 1981 to 1986. The largest single yearly decline to date was posted after the October 1983 - September 1984, production period. By contrast, after the October 1984 to September 1985 production of 121.4 million gallons, the BLM well recovered essentially to its previous year's level.

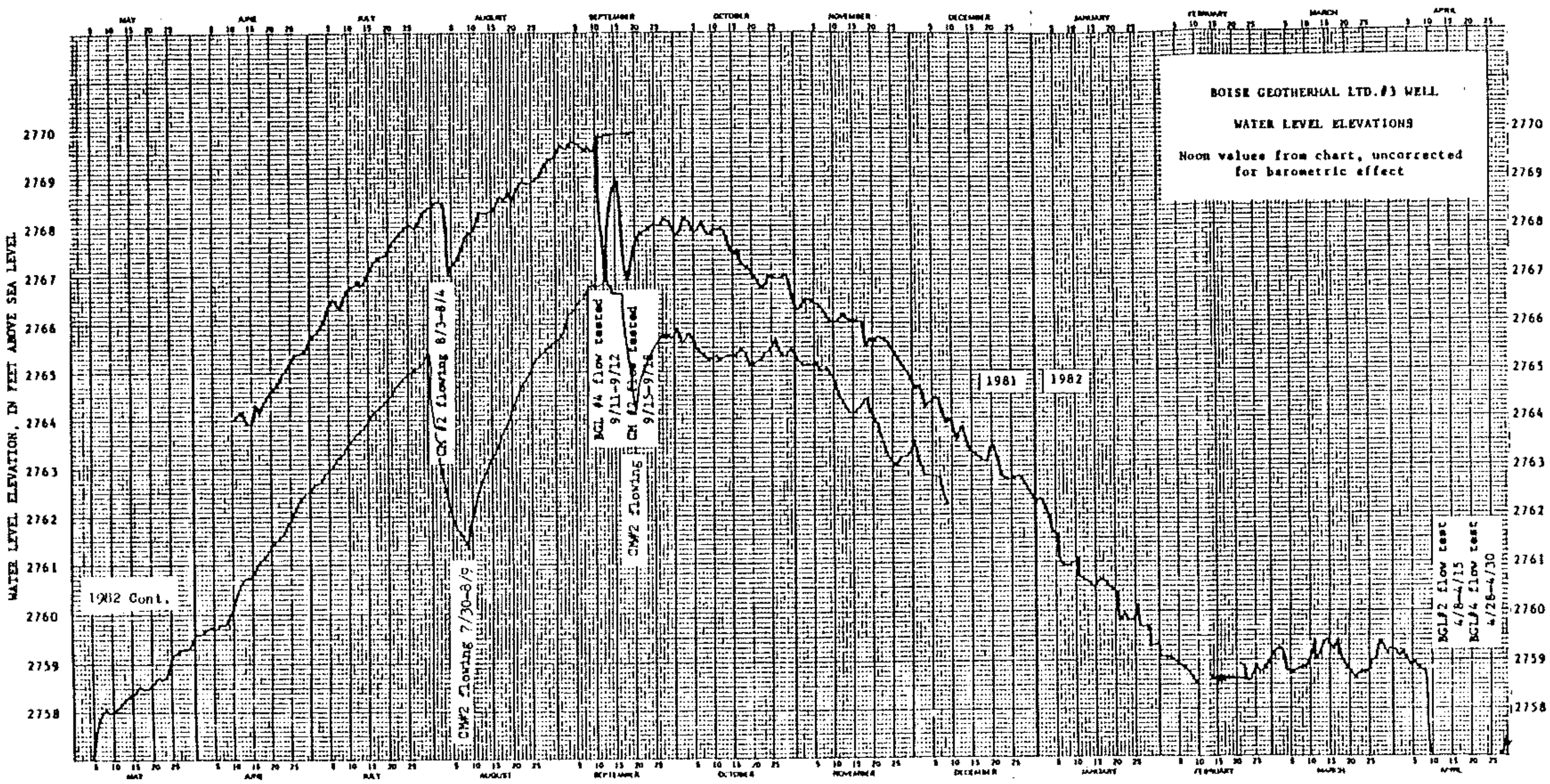
The latest production figures provided by the City of Boise indicate that BGL extractions during the period October 1, 1985 to September 30, 1986 total 176.8 million gals. It has already been mentioned that recovery in the BLM well peaked at approximately 2,745 feet in September 1986, posting a 4.5 foot decline. Surely other factors in addition to BGL pumpage influence this part of the geothermal system. Some factors cannot be identified with certainty, others cannot be adequately evaluated at this point in the study because the data base is not long enough. However, if those factors continue to affect the potentiometric surface in the manner which they have since BGL came on line in October of 1983, then it would appear from the drawdown - recovery and production relationship that withdrawals in excess of 125 million gallons will accelerate the antecedent decline in the potentiometric surface in the vicinity of the BLM well.

Targeting and evaluating those "other factors" that are causing annual fluctuations and the decline in the potentiometric surface in the vicinity of the BLM well is difficult, at least at this stage of understanding of the system. Logically, all of the annual fluctuations in the BLM hydrograph since October 1983 cannot be attributed to BGL pumpage because Figure 10 indicates that more subdued annual fluctuations on the same seasonal cycle were evident in the well at least as far back as 1981.

CAPITOL MALL - BLM HYDROGRAPH FLUCTUATION

Figure 12 is a hydrograph from the BGL No. 3 well, and was adopted directly from Burnham and Wood (1983). The BGL hydrograph also shows an annual fluctuation in potentiometric surface similar to that in the BLM well. In addition, the curves show testing periods for the Capitol Mall No. 2 well (CM No. 2). The abrupt declines in water-level in BGL No. 3 when CM No. 2

FIGURE 12



*ADOPTED FROM BURHAM & WOOD, 1983

is allowed to flow indicates rather good interconnection between the two wells. Flow tests on BGL No. 4 also show abrupt declines in the BGL No. 3 hydrograph indicating good interconnection between BGL No. 3 and No. 4. It has already been established that good interconnection exists between the BGL production wells and the BLM well; thus, it follows that Capitol Mall No. 2 well production should have a direct effect upon the potentiometric surface at the BLM Well. Therefore, since 1982, when Capitol Mall No. 2 began production, the effects of that production are probably manifested in the water-levels in the BLM well. It follows then, that the fluctuations and cumulative declines in recovery levels noted in September of recent years including 1983, 1984, 1985, and 1986 (Fig. 10), are partly due to Capitol Mall withdrawals. The effects of Capitol Mall production of geothermal water on the BLM well, however, are likely to be small compared to BGL production effects. CM No. 2 is much further away from the BLM well than the BGL wells (Fig. 1). Capitol Mall No. 2 production is also reinjected at CM No. 1. Although the details of the aquifer are not yet understood, such injection would serve to at least partially repressurize that portion of the aquifer system. Since good interconnection between the BGL, Capitol Mall, and BLM wells evidently exists, that portion of the aquifer will be referred to as the BGL-BLM-CM portion.

BLM HYDROGRAPH FLUCTUATION NOT RELATED TO BOISE GEOTHERMAL LIMITED OR CAPITOL MALL PUMPAGE

In addition to the BGL and CM effects, still other factors which yield the annual September to September hydrograph fluctuation are evident in Figure 10. During the period September 1981 to September 1982, neither Capitol Mall nor Boise Geothermal Limited Wells were producing from the aquifer. Yet, as noted, an annual fluctuation is evident.

In attempting to explain similar cyclical annual fluctuations in the BLM water level during 1976-1978, Nelson and others, (1980, p.23) state, "An obvious conclusion would be to suspect the water level drops that occur after December of each year are the result of increased water usage by the Warm Springs wells for the winter heating months." However. they finally preferred to explain these annual cyclical fluctuations as a response to loading and unloading of the confined aquifer and its confining layer by the cold water reservoir. The increases in water-level in the cold water reservoir resulting from the influx of irrigation waters beginning in the spring, continuing throughout the summer and declining after irrigation ceases in the fall, create a cyclical load on the geothermal system. Such an effect would be similar to tidal effects in coastal confined aquifers where the weight of the incoming tidal waters load and compress the confined aquifer to some extent and cause an increase in water levels in wells penetrating the confined aquifer. The water-level rise in a well, divided by the water-level rise in the tide is termed the aquifer's tidal efficiency (TE) and is a reflection of the rigidity or elasticity of the aquifer.

Fracture system aquifers such as the Boise Geothermal system, commonly have a high rigidity and a correspondingly low tidal efficiency. Tidal efficiency (TE) is also related to barometric efficiency (BE) by the relationship TE + BE = 1 (Ferris, et.al., 1962). Preliminary calculations of the barometric efficiency of the aquifer in the vicinity of the Kanta well indicate an efficiency on the order of .8 or .95 or 80-95%. This value compares favorably with a published barometric efficiency of 90% for BGL well No. 3 (Kelly, J.E., 1986). If these values are accepted as approximating an average for the entire system, then the tidal efficiency of the aquifer would be approximately 10% or .1.

Figure 10 shows a water-level fluctuation of approximately 10-11 feet between September 1981 and mid-February 1982. Similar fluctuations on the same annual cycle were recorded by Nelson and others (1980) during 1976, '77, and '78. Using a tidal efficiency of 10% and assuming no cold water leakage in the geothermal aquifer, a cold-water table fluctuation of approximately 100-110 feet would be required to induce the potentiometric surface fluctuations recorded in the geothermal aquifer between September and late February or early March. Nelson and others (1980, p. 24) for the years of record 1934-77, show a maximum range of annual fluctuation of approximately 10 feet for the free water table near Meridian. This is approximately 1/10 of the annual head change that would be required to cause the fluctuations observed in the geothermal system. Or, conversely, the geothermal aquifer and all of the overlying confining layers would have to have a tidal efficiency of about 100%, perfect elasticity, for the phreatic reservoir to provide the required load changes.

The tidal concept is generally attractive theoretically and the timing of loading and unloading appears to be close enough. However, the indication of the rigid nature of the geothermal aquifer requires cold water-table fluctuations which are much larger than generally observed. Perhaps such loading and unloading contributes to the observed fluctuation, and is integrated with other causes, but it does not seem likely, at least at this junction in the study, to be the sole cause.

The cyclicity of fluctuations may also reflect, to some degree, annual changes in the balance of recharge and natural discharge within the system. Nelson and others (1980, p. 23) also suggest that the annual geothermal fluctuation in the BLM well may be caused by vertical leakage from the cold water aquifer. Admittedly, little is known about how and where the geothermal system becomes recharged. However, if differences in the rate of recharge to the system are to be called upon to explain the fluctuations, then it should be recognized that it is likely that the principal recharge to the geothermal system takes place high in the batholith and not by vertical leakage as described by Nelson. More likely, warm waters from the geothermal aquifer leak upward into the cold water aquifers because of the higher artesian pressures observed within the geothermal aquifer. Indeed, this may be at least partly the explanation for the anomalously high temperatures found in some of the deeper "cold" water aquifers.

BOISE WARM SPRINGS WATER DISTRICT PUMPAGE RELATED TO BLM HYDROGRAPH(?)

Having the benefit of additional data from subsequent years, and knowledge gained from monitoring the Warm Springs portion of the system for the past two years, it seems more likely that all or certainly the largest part of those cyclical fluctuations in the BLM well are, indeed, due to the influence of pumpage by the BWSWD wells 1 and 2. The Warm Springs wells were the only wells making very large withdrawals from the geothermal system until the winter of 1982. Complete pumpage data are available for the period from 9/1/81 to 3/1/82, the decline period, when approximately 190 million gallons were withdrawn by the Warm Springs wells. Although the aquifer system is apparently faulted into blocks, it seems unlikely that an essentially impermeable barrier or series of impermeable barriers separate the Warm Springs portion of the system from the BLM-BGL-CM portion. Thus, it seems quite probable that large drawdowns for prolonged periods such as those created by Warm Springs pumpage could cause or at least contribute significantly to the measured fluctuations in the BLM well.

Unfortunately, estimates of the possible drawdown effects that one might expect as a result of Warm Springs pumpage are difficult to make from the data. Nevertheless, using the available information and by making what seem to be reasonable assumptions one can suggest, at least, a range of drawdown effects. Such estimates require values of transmissivity (T) and storativity (S) of the aquifer. Because the aquifer is a fracture-controlled system, T & S values are difficult to determine and are variable throughout the aquifer depending largely on the abundance, openness, and continuity of the fractures.

The mathematical models applied in this estimate are designed for a porous media aquifer rather than fracture systems. However, all presently available transmissivity and storativity estimates for the aquifer were also made using porous media models. Wood and Burnham (1986) report on data gathered from discharge-drawdown tests of several types, from artesian head response to barometric pressure change, and from long-term response of artesian head to variable annual discharge rates. Their analyses have been strictly for engineering-design purposes. In spite of that, they suggest that within time limits of a few hours to a few days, the system responds to a measured discharge-drawdown stress as though the average transmissivity is of the order of 240,000 gal/day/ft (32,000 ft^2/day, or 3,000 m^2/day) and storativity is of the order of 5 x 10 EXP -4.

Although these values have been useful for design of wells, they should not be applied to the aquifer system as a whole.

Waag and Wood (1985) published estimates of transmissivity in the vicinity of the BWSWD and Kanta wells of 3500 gal/day/ft. 13,000 gal/day/ft and 25,000 gal/day/ft. The highest of the estimates was based on the longest term sample. Recognizing that the limiting assumptions for strict use of the porous media mathematical models cannot be met in the Boise Geothermal System, it, nevertheless, seems worthwhile to make a few qualified estimates of the possible drawdowns that might be expected in the BLM well as a result of pumpage by Warm Springs wells.

Considering then the period 9/1/81 to 3/1/82 when the BLM hydrograph in Figure 10 shows a 10-11 ft. decline and when BWSWD pumpage was 190 million gallons, averaging 730 gpm, and using the following forms of the Theis equation:

$$u = \frac{1.87\ r^2\ S}{Tt} \quad \text{and} \quad s = \frac{114.6\ Q\ \ W(u)}{T} \quad \text{where}$$

W(u) = Exponential integral

r = Distance in feet, from the discharging well to the point of observation

S = Storativity

T = Transmissivity

t = Time in days since pumping started

s = Drawdown, in feet, at any point of observation (r) in the vicinity of a well discharging at a constant rate

Using values of:

$T = 25{,}000$ gal/day/ft and
$S = 5x10^4$
$Q = 730$ gpm
$r = 7580$ feet (approximate map distance from BSWD No. 2 to the BLM well)
$t = 181$ days - 9/1/81 to 3/1/82 and obtaining values of the well functions of u (Wu) from standard tables (Ferris et.al., 1962, p.96-97)

a value of s = 12.9 ft. is obtained.

Using a $T = 240{,}000$ gal/day/ft

$S = 5x10^{-4}$

and other values as above, the estimated drawdown decreases to 2.1 ft.

These estimated drawdown values between 12.9 and 2.1 feet are probably reasonable for the range of effects that might be expected at the radius of the BLM well owing to BWSWD pumpage for the 181 day period. The transmissivity value of 240,000 gpd was obtained from pump testing in the BGL portion of the system. Quite likely it is valid for that vicinity, but it is nearly an order of magnitude greater than our largest estimate of T in the Warm Springs portion of the aquifer using longer term, but not as well controlled, pumpage data. A transmissivity of 240,000 gal/day/ft is probably too large to apply to the whole portion of the aquifer system between the BWSWD wells and the BLM well. Thus, if the BLM well does, indeed, experience drawdown as a result of pumping by the Warm Springs wells the water level declines would be expected to exceed 2.1 feet. Lower values of transmissivity would shift the estimated drawdown toward the upper side of the range of 12.9 feet calculated above, perhaps, to drawdown values similar to those shown in Figure 10 for the period 9/1/81 to 3/1/82.

DECLINING POTENTIOMETRIC SURFACES IN THE WARM SPRINGS AREA

Perhaps another evidence for interconnection between the Boise Geothermal Limited - BLM - Capitol Mall portion and the Warm Springs portion of the geothermal aquifer system are the generally declining water levels within both portions of the aquifer system. The declining water levels in the BGL-BLM-CM portion, although still rather small, should be viewed with interest and concern. They reinforce the need for further documentation and understanding of the total system, especially when the BGL-BLM-CM portion of the system is considered in the context and timing of more severe declines evident in the Warm Springs area. According to BWSWD records, the last time that artesian surface flow occurred at the pumping wells was in early August 1983. During that summer, well No. 1 flowed for 14 days and well No. 2 experienced very small flows for 12 days. Since at least 1983, the water-level in the pumping wells has been in a general decline. Maximum recovery during the summer of 1984 came only to within 10 feet of the surface. During the summer of 1985, the recovery levels hovered between 20 and 25 feet below the surface; except on the 29th of August when for one day it reached to within 15 feet of the surface. By comparison, prior to 1983 surface flows were common during the summer in that portion of geothermal system. An accurate analysis of the flow at the wells during recovery is not possible because many of the older records from the summer period are quite sketchy and difficult to interpret. However, the recovery level and flow data summarized in Table II lend some impression of the duration of flow as recorded.

TABLE III

Maximum Recovery of Potentiometric Surface BWSWD Wells 1 and 2

1986	25 feet below surface
1985	15 feet below surface
1984	10 feet below surface
1983	At surface - Flow 14 days
1982	At surface - Flow 7 days
1981	At surface - Flow 50 days
1980	At surface - Flow 36 days
1979	At surface - Flow 38 days
1978	At surface - Flow 6 days

As might be expected, the declines in the maximum recovery levels of the potentiometric surface have been attended by increases in the depths of the pumping levels with the Warm Springs wells in recent years. Pumping level as used here refers to the water level during pumping and not the depth at which the pumps are installed.

Table III shows the various depths to water-level (pumping level) averaged for well No. 2 for the period June 15-June 15 for the years 1978 through 1986. According to the data in the table, an abrupt decline in the average pumping levels in well No. 2 occurred between June 1983 and June 1984. Coincidentally this includes the period during which the BGL wells had their large initial production, and in which the BLM well registered its largest drawdown and subsequent decline in recovery level. If this is not just coincidence, then it appears to be another indication that the Warm Springs and the BGL-BLM-Capitol Mall portions of the geothermal system are, indeed, interconnected and that sustained production in one portion of the system manifests drawdown in the other portion.

TABLE IV

YEARLY AVERAGE DEPTH TO WATER LEVEL
BWSWD WELL NO. 2
(pumping level)
June 15 to June 15

YEAR	WELL NO. 2
1985-1986	103.4
1984-1985	105.2
1983-1984	103.2
1982-1983	81
1981-1982	80.8
1980-1981	65.4
1979-1980	71.2
1978-1979	102
1977-1978	107

Clearly there is a need for further monitoring and analysis of the system before such a relationship can be firmly established. A data base long enough for a comprehensive study of the system is just now becoming available. Compared to using long-term monitoring of the system to identify interconnection, it should be considered that drawdowns created in one portion of the system by flow or pump testing in other portions for relatively shorter periods of time may be too subtle and/or too overprinted by other effects to be recognized. For example, in April 1982, BGL well No. 2 was flow tested at 900 gpm for seven days and BWSWD well No. 3 was used as an observation well. Apparently the flow tests did not produce identifiable effects in BWSWD well No. 3. This is understandable. If one uses the same Transmissivity and Storativity values, assumptions, and Theis equations used earlier in this report to estimate the effects of BWSWD production on the BLM well, the estimated drawdown created in the No. 3 well by the flow tests performed in BGL No. 2 would range from .5 to 3.7 feet. One would normally expect to be able to recognize drawdowns within that range. However, review of records from Warm Springs wells 1 and 2 indicates that water-level recoveries of 75 and 50 feet respectively occurred during the test period. The distance between the BWSWD pumping wells, which were producing during the test period, and No. 3 is only 645 feet. Considering that all three Warm Springs wells apparently enter the same fracture zone,it is quite likely that any drawdown in No. 3 caused by the flow testing of BGL No. 2

would be off-set and overprinted by the large recovery in the pumping wells and the local fracture network.

To see if the recent lowering of average pumping levels and declining recovery levels in the Warm Springs area are simply related to increased extractions by the Warm Springs system itself, production was estimated. Monthly and annual production data for the Boise Warm Springs Water District are in Table IV. Although some gaps in the data exist, it is apparent that production by the District has not increased since the fall of 1983. Indeed, the data indicate that the District has experienced a decline in production since 1980. In fact, in recent years wells 1 and 2 have had to pump from greater average depths to obtain equal or smaller quantities of water.

Review of the longer term production and drawdown data from both portions of the system, suggests that a pattern of evidence indicating interconnection and interference is emerging. It must be emphasized again, though, that it is obvious from the data that other factors also influence recovery and pumping levels within the system. For instance, during the period June 15, 1978-1979 and 1977-1978, the pumping levels in well No. 2 averaged 102 feet and 107 feet respectively. These are pumping depths equal to or exceeding those experienced since 1983. Those periods were, however, succeeded by recoveries which caused artesian flow; phenomena which have not been observed in the Warm Springs wells since the summer of 1983. It appears that in addition to the obvious annual cycles, other multi-year cycles exist. The available data base is still too short to evaluate that possibility. Nevertheless, whether a correlation exists between regional drought and water levels within the geothermal aquifer is an important consideration. For instance, what is the role of precipitation and snow accumulation, if any, in recharge to the aquifer? Such a study is the next logical step in a comprehensive study of the system.

TABLE V

PRODUCTION
BOISE WARM SPRINGS WATER DISTRICT

PERIOD	MONTHLY PRODUCTION (M GALLONS)	ANNUAL SEPT. 1 - AUG 31. (M GALLONS)	ANNUAL SEPT. 1 - AUG 31 (ACRE FEET)
April '78	21,927		
May	14,753		
June	9,577		
July	5,838		
August '78	3,366		
September '78	3,717		
October	5,799		
November	31,948		
December	40,740		
January '79	48,331		
February	49,295		
March	42,799		
April	29,587		
May	19,653		
June	12,778		
July	8,751		
August '79	9,321	1978-1979 302,719	922.5
September '79	11,199		
October	21,405		
November	36,143		
December	39,193		
January '80	49,230		
February	39,842		
March	35,372		
April	22,306		
May	18,779		
June	15,101		
July	7,421		
August '80	9,109	1979-1980 305,100	935.8

PERIOD	MONTHLY PRODUCTION (M GALLONS	ANNUAL SEPT. 1 - AUG 31. (M GALLONS)	ANNUAL SEPT. 1 - AUG 31 (ACRE FEET)
September '80	14,140		
October	22,678		
November	30,418		
December	36,151		
January '81	37,241		
February	32,811		
March	20,315		
April	13,782		
May	10,324		
June	5,868		
July	6,200		
August '81	6,634	1980-1981 236,562	725.7
September '81	13,151		
October	26,008		
November	30,548		
December	35,667		
January '82	46,778		
February	38,050		
March	------		
April	------		
May	------		
June	------		
July	------		
August '82	5,892	1981-1982 196,094	601.5
September '82	13,288		
October	23,612		
November	35,496		
December	40,744		
January '83	37,122		
February	29,038		
March	30,059		
April	26,641		
May	19,530		
June	10,038		
July	9,269		
August '83	7,048	1982-1983 281,885	864.7

PERIOD	PRODUCTION (M GALLONS)	ANNUAL SEPT. 1 - AUG. 31 (M GALLONS)	ANNUAL SEPT. 1 - AUG. 31 (ACRE FEET)
September '83	15,890		
October	24,859		
November	31,763		
December	41,336		
January '84	41,438		
February	34,021		
March	32,908		
April	28,796		
May	19,534		
June	8,392		
July	------		
August '84	------	1983-1984 279,937	858.7
September '84	------		
October	29,467		
November	31,250		
December	36,842		
January '85	37,557		
February	32,714		
March	34,055		
April	16,367		
May	------		
June	------		
July	------		
August '85	8,791	1984-1985 227,043	696.5
September '85	21,264		
October	29,314		
November	36,322		
December	36,808		
January '86	32,882		
February	25,310		
March	23,140		
April	20,150		
May	15,507		
June	6,638		
July	6,893		
August '86	5,628	1985-1986 259,856	797.1

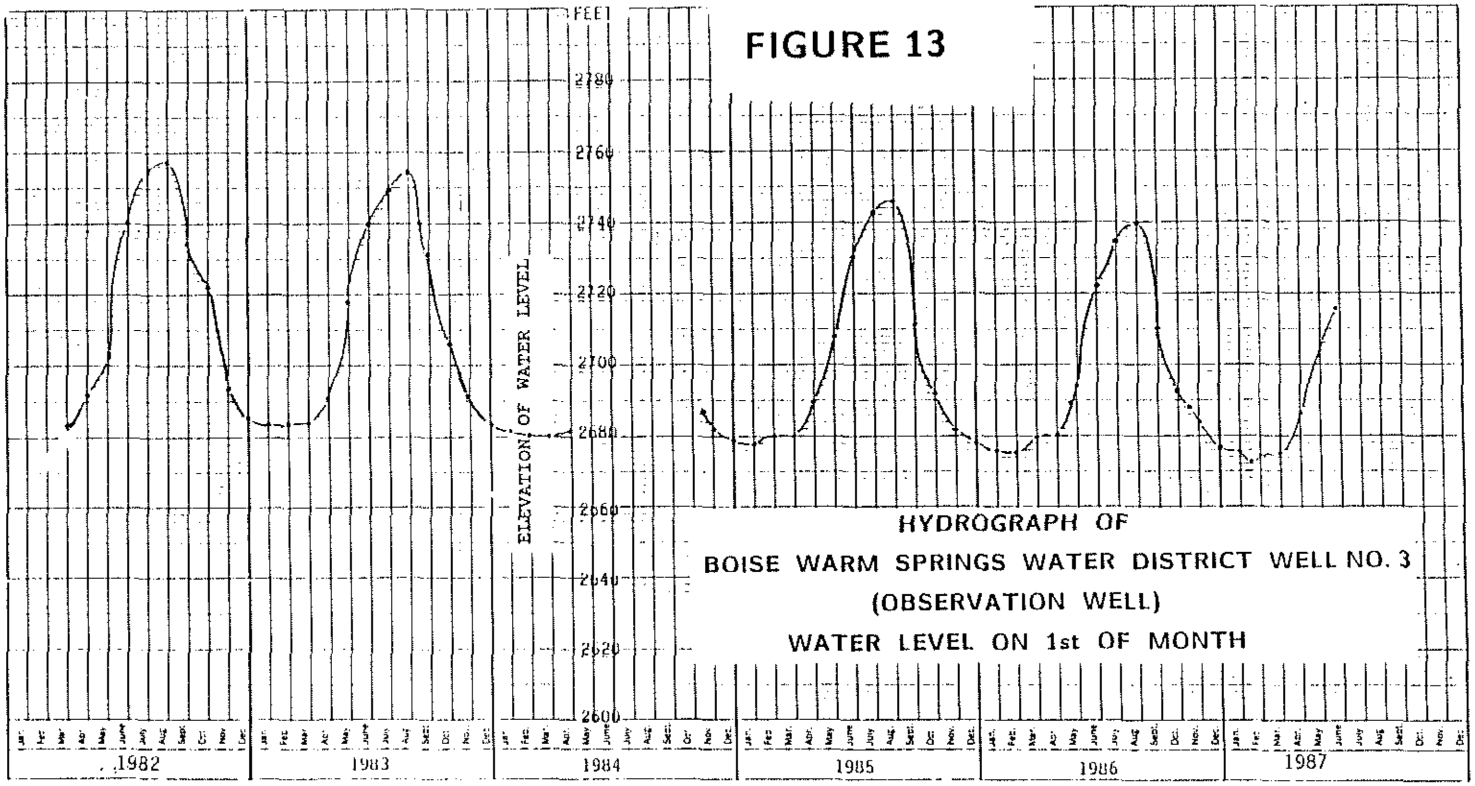
DECLINES IN THE BWSWD WELL NO 3

Other wells in the vicinity of the Warm Springs pumping wells also show signs of decline in recent years. Data from Warm Springs Water District well No. 3 are shown in Figure 13. It is also informative to note that the maximum recovery in No. 3 during August of 1982, reached 2758.5 feet, 18.5 feet above its 1986 recovery. That is an average decline of 4.6 feet per year. Some caution must be expressed concerning attaching too much significance to a few feet of difference in the recovery levels in well No. 3 on a short-term basis. Because the interconnection between No. 3 and the pumping wells is well developed, No. 3 responds quickly to pumpage and may fluctuate several feet on a short term. However, longer term declines in maximum recovery similar to that evidenced since 1981 cannot be explained by short-term pumping effects.

DECLINES IN THE KANTA WELL

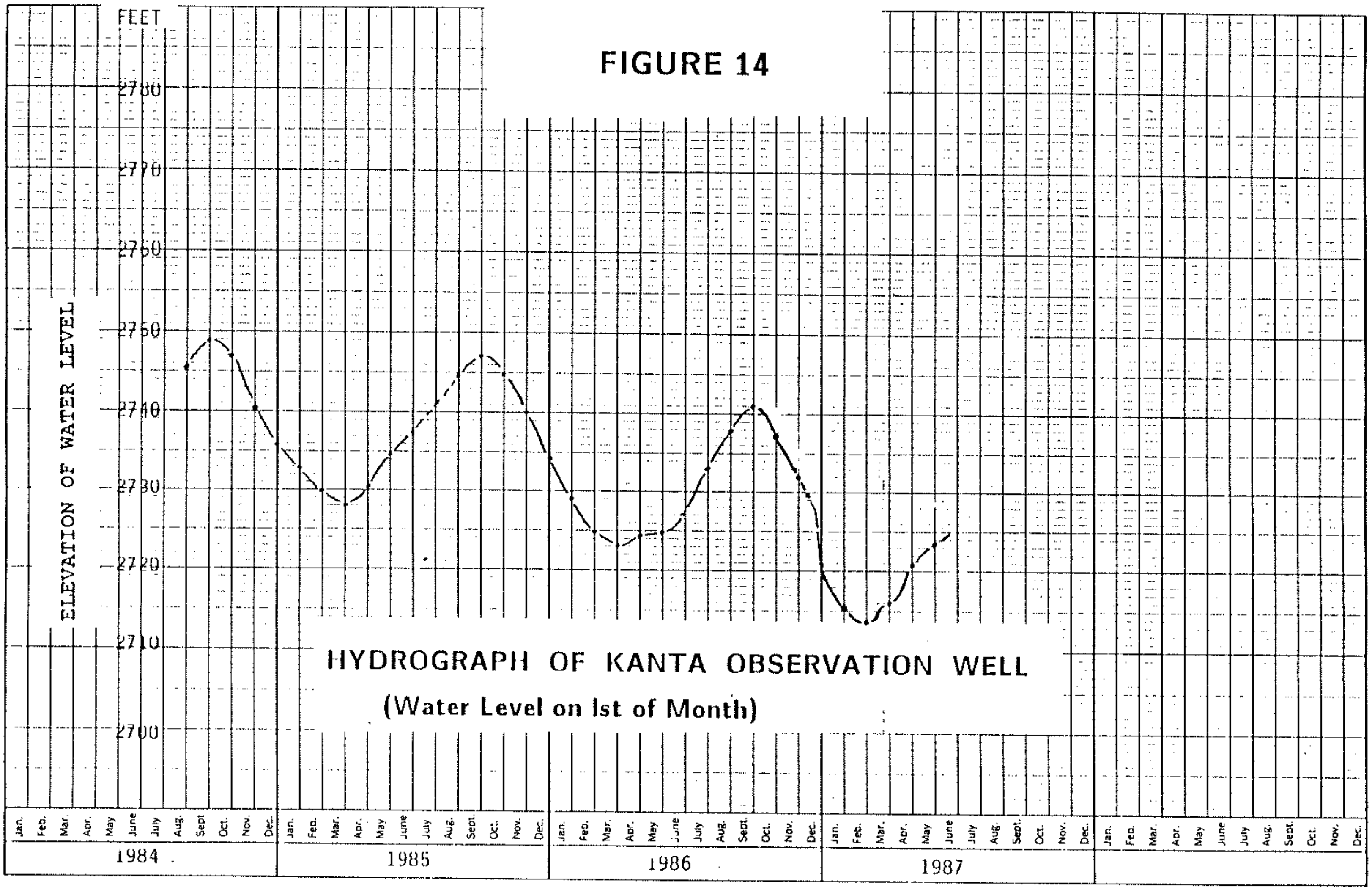
Figure 14 is a hydrograph of the Kanta well. Although the period of record is short, it clearly shows an increase in the maximum drawdown and an accelerated decline in the maximum recovery within the past two years. The Kanta well reflects the general trend within the aquifer system, but the situation is probably exacerbated by extractions by the State of Idaho Penitentiary well which is approximately 675 feet, S 10° E of the Kanta well. In the spring and summer of 1985 and 1986, the State

FIGURE 13



HYDROGRAPH OF
BOISE WARM SPRINGS WATER DISTRICT WELL NO. 3
(OBSERVATION WELL)
WATER LEVEL ON 1st OF MONTH

FIGURE 14


FEET
2780
2770
2760
2750
2740
2730
2720
2710
2700
ELEVATION OF WATER LEVEL
HYDROGRAPH OF KANTA OBSERVATION WELL
(Water Level on Ist of Month)
Jan. Feb. Mar. Apr. May June July Aug. Sept Oct. Nov. Dec.
1984
1985
1986
1987

well was pumped to provide irrigation water for the gardens of the Idaho Botanical Garden Society. The amount of pumpage from the well has not been measured, and no attempt has been made to determine what effects, if any, the State well production might have upon the Kanta well and other wells in the system.

As a part of the 1984-85 IWRRI funded study of the Warm Springs portion of the aquifer, a transmissivity value of 6800 gals/day/ft for the aquifer in the vicinity of the Kanta well was estimated (Waag and Wood, 1985). This value is considerably less than the estimate of approximately 25,000 gals/day/ft. for the transmissivity of the fracture system in which BWSWD wells 1 and 2 are completed (Waag and Wood, 1985). The Kanta well responds well to pumpage by BWSWD Wells 1 and 2 indicating good interconnection. However, the lower transmissivity and knowledge of the structure in the area prompts the interpretation that the Kanta well is drilled in a down-thrown block of the fracture system in which wells 1 and 2 are completed. Although exact correlation of individual fluctuations in the hydrographs of BWSWD wells 1 and 2 (Figs. 4, 5) to well No. 3 and the Kanta well (Figs. 6, 7) is uncertain the general similarity of hydrographs indicates that good interconnection between the wells exists.

DECLINES IN THE QUARRY VIEW PARK WELL

In addition to extractions by the State well, the city owned well at Quarry View Park has also produced water for irrigation from the geothermal system. From late April until mid-October 1985, the Quarry View well pumped an estimated 15.8 million gallons. During the period May through September 1986 an additional 13.9 million gallons were produced. These figures should be regarded as "ball park" estimates only. The well is not equipped with a flow meter and these estimates are based upon electrical power consumption using one of the standard formulas (Young et.al., 1979). Appendix A. Proper application of the formula requires a knowledge of the drawdown during pumpage. All measurements during this study were taken when the well was not pumping; thus, some amount of recovery had already occurred.

To estimate the production, an average water level during pumping of 135 feet was used. This depth is based upon water level measurements taken during this study and drawdowns experienced during aquifer tests performed on the well by Anderson and Kelly in February of 1984. The test data when analyzed using the Jacob Straight-line mathematical model for porous aquifer systems yield a transmissivity of approximately 1000 gal/day/ft. This value is considerably lower than the transmissivities estimated for the aquifer in the vicinities of the Warm Springs wells and the Kanta well (Waag, and Wood, 1985).

It is difficult to judge from the drillers log; however, the well appears to penetrate only the upper 100-150 feet of the geothermal aquifer. Perhaps, the large drawdowns experienced and the low transmissivity indicated by the test data are more reflective of this partial penetration and completion problems with the well rather than being representative of the aquifer in the vicinity of the well. At this stage of the study data to determine how or if the Quarry View well responds to pumpage by the Warm Springs wells is not available. It is worthy of note, however, that although the Quarry View well is only about one half as distant from the pumping wells, it is not as responsive as the Kanta well. This suggests that the Kanta and Quarry View wells are in segments of aquifer that have different characteristics and are, to some degree, separated from each other and from the Warm Springs fracture zone.

DECLINES IN THE BEHRMAN WELL

The Behrman well is remarkable in that its fluctuation appears to be related to Warm Springs pumpage, but it is slow to respond and the response is very attenuated. Whereas the pumping wells sustained drawdowns of 135 feet and BWSWD well No. 3 a drawdown of 71.2 ft., the Behrman well registered a maximum fluctuation of approximately 20 feet. Yet as shown in Figure 1, the Behrman well is only approximately 85 feet further from the pumping wells than well No. 3. The lowest water level measured in the Behrman well during the monitoring period was on February 25, 1986 at 3 feet below the well head, or at ground elevation (2735.7 feet). The lowest water level in BWSWD well

No. 3 was also recorded on February 25 at 113 feet below the collar (2789.5) at an elevation of 2676.5 feet. This reflects a head difference of 59.2 feet between the Behrman well and well No. 3. Such a notable head difference between these two non-pumping wells which are only about 320 feet apart suggests that a boundary exists between the Behrman well and Boise Warm Springs Water District wells, 1, 2, and 3. The nature of the barrier, however, is not known. Unfortunately, no geologic or drill hole information on the Behrman well is available. It may be that the Behrman well is simply drilled into the downthrown block of the fracture zone in which BWSWD wells 1, 2, and 3 are sited. In that case, it must be a different block than the one in which the Kanta well is drilled. It will be recalled that the Kanta well is much more responsive to pumpage by wells No. 1 and 2, yet the Kanta well is not completed within the BWSWD fracture zone either.

Whatever the character of the aquifer block in which the Behrman well lies, the recovery level in the well reached its maximum 1986 recovery on September 1 at 7.5 psi. This is equivalent to a head of 17.3 feet of water above the gauge height or 20 feet above ground level and an elevation of 2755 feet. Interestingly, this level of recovery exceeds, by approximately 10 feet, the 1986 maximum recovery levels of the water in all of the other wells monitored during this study. Part of its anomalous drawdown and recovery behavior may well be attributed to well characters. It is, however, difficult to dismiss all of its anomalous behavior as owing to well completion, and differences in aquifer characteristics in the vicinity of the well. Thus its performance compared to the Warm Springs, Kanta, and Quarry View wells deserve further consideration and study.

Current observations suggest that the Behrman well's relationship to other wells in the system is complex. It seems to be separated from the Warm Springs wells and the Kanta well by a semi-permeable barrier or barriers. Its degree of interconnection with the Quarry View and the State well are even less well understood and a more comprehensive and longer data base will be required to understand these relationships.

CONCEPTUAL MODEL OF THE GEOTHERMAL SYSTEM

Several models may be developed to characterize both the heat supply and the areal geothermal groundwater circulation system. The one now thought most likely is a system of circulation to about 1 mi (2 km) depth over a path of about 6 mi (10 km) through north-to-northeast-trending deep fracture zones in the Idaho batholith, Fig. 15 (Wood and Burnham, 1983, 1987). Meteoric-water recharge at high elevations circulates with an indicated residence time (Carbon-14 activity) of 6,700 to 17,000 years (Mayo and others, 1984). Discharge at the western Snake River Plain margin is through seeps, springs, or by lateral migration through fractured silicic volcanic rock. The regional high heat flow anomaly (>2.5 microcalories/cm^{2}s of the northern Basin and Range (Lachenbruch and Sass, 1978) apparently extends to the western Snake River Plain and the southern part of the Idaho batholith. Geothermal gradient in the plain and in the region of the geothermal wells is more than 40^{o}C/km (wood, 1984), which provides opportunity for the observed well-discharge temperatures with relatively shallow circulation. Further, heat flow along the margins of the western Snake River Plain is typically 3.0 microcalories/cm^{2}s, possibly localized by thermal refraction related to the faulted interface of relatively conductive granitic rocks of the margin with the more insulating basin fill deposits of the plain (Brott and others, 1978).

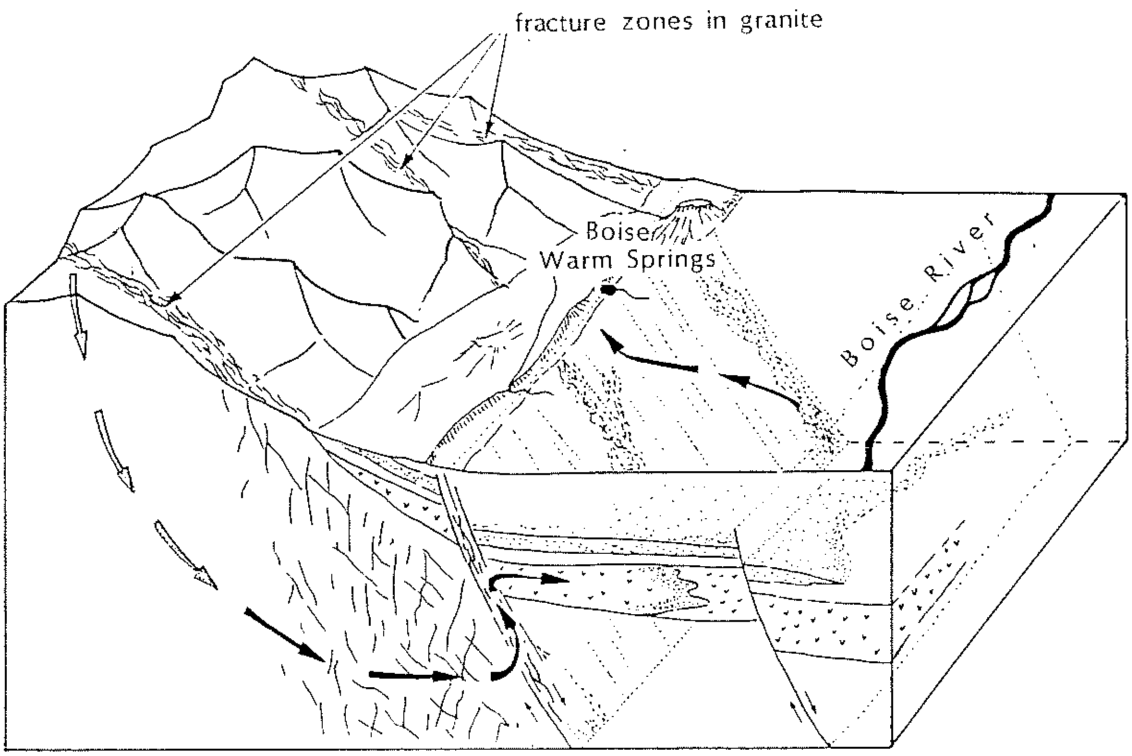


Figure 15. Conceptual model of the geothermal groundwater circulation system through fractured granite to the discharge area along the foothills fault zone of Boise and into the permeable rhyolite aquifers beneath the north-eastern part of the city. (Adopted from Wood and Burnham, 1987)

CONCLUSIONS & SUGGESTIONS FOR FURTHER STUDY

Although the data base is short and limited to only portions of the geothermal system, enough data has been accumulated to identify trends within the aquifer and to obtain a preliminary understanding of some of the inter-relationships between segments of the aquifer system. The data indicate that production from the Warm Springs and BLM-BGL-CM portions of the geothermal system exceeds the ability of those portions of the system to recover on an annual basis. The peak recovery in the BLM well in 1982 reached an elevation of approximately 2,762 feet. The four successive recovery peaks since 1982 show a persistent downward trend declining to an elevation of 2,745 feet in 1986. In 1981 recovery reached 2,763 feet in the BLM well. Prior to 1981, data are somewhat spotty but recorded observations by Nelson, et.al. (1980) indicate a recovery to 2,761[6] feet in September 1978, 2,758 feet in mid-September of 1977 and 2,758 feet as late as November 1, 1976. It may be assumed that the recovery peak, during fall of 1976, was higher since all records show some decline after early September. At any rate, these data suggest that the maximum recovery of the potentiometric surface in the BLM-BGL-CM portion of the aquifer prior to 1983 was fairly consistent reaching approximately 2,760 feet in late August or early September indicating that portion of the system was in equilibrium or near equilibrium conditions.

The Warm Springs portion of the geothermal system has also experienced declines in maximum recovery levels since 1983. The recovery of 1983, the summer before BGL began production, was the last time that BWSWD wells 1 and 2 experienced flow at the surface. Since that time, maximum recovery levels in those Warm Springs wells has declined 20-25 feet. Concomitantly, average yearly pumping levels in the principal pumping well, BWSWD No. 2, have declined from 83 feet to 103 feet. These decreasing recovery levels and increasing pumping depths apparently are not simply the result of increased BWSWD production. In fact, the data indicate that there has been a small decline in annual production since 1978. Thus, the Boise Warm Springs Water District is pumping from greater average depths to obtain the same or less production. Because the recovery declines in the BWSWD and the BLM-BGL-CM segments of the aquifer have accelerated since BGL production began, the probability that the two segments are hydraulically interconnected must be given serious consideration. Furthermore, although the evidence is not unequivocal, the most probable explanation for the annual fluctuations in the BLM-VA hydrographs prior to 1982 is that those fluctuations are principally the manifestation or influence of pumpage by Boise Warm Springs Water District wells 1 and 2. If that is so, then the nearly flat recovery levels in 1976, '77, '78, '81, and '82 in the BLM well 1 not only indicate equilibrium conditions with the aquifer but interconnection. Thus, on the basis of the data it is more realistic, accurate, and prudent to view the BWSWD and BLM-BGL-CM portions of the system as integral parts of a single aquifer which is segmented geologically by fractures, but hydraulically interconnected.

To improve, expand, and refine understanding of the geothermal system, the data collection network now in place should be continued to extend the data base. The monitoring network should be expanded to include the Edwards well and the Milstead well, the Harris well, and one or more of the BGL wells. Production, pressure, and temperature data from the Capitol Mall wells are presently available and should be plotted and analyzed for trends. Reliable temperature data from the larger production wells are lacking and a comprehensive program of temperature data collection over the drawdown and recovery cycle should be started. A scattering of data on pH and water chemistry is presently available, but a systematic data base of these parameters is not. As noted, some data available from various wells within the system show differences in pH and ion concentrations. However, the data were not systematically obtained, thus, they are difficult to interpret. It is not known, for instance, if the temperature, pH, and chemistry of the wells within the system change with prolonged pumpage during the production season. It may be that if trends exist, they will contribute to understanding flow patterns, recharge, and rock types of the aquifer system. The Dallas Harris well, which has recently been deepened to approximately 690 feet, is terminated in granite. The principal production wells in the geothermal system produce from the rhyolite aquifer or from a granite-rhyolite fracture zone. A complete inorganic chemical analysis of the Harris geothermal water is needed to determine if the differences in granite versus rhyolite geothermal waters noted in the chemistry section of this report can be corroborated within the known boundaries of the system. To estimate its mean subsurface residence time in the fractured granite aquifer, a radiocarbon date of the water from the Harris well should also be obtained.

The water-level declines evident in the BWSWD and BLM-BGL-CM portions of the aquifer system indicate that the system is under stress. Demand on the system is increasing and greater declines are anticipated. BGL is currently expanding production and plans even greater expansions in the near future. The water-level data currently on hand show no evidence of approaching equilibrium and expansion of production will certainly postpone equilibrium if it is in the offing. In late 1987 or early 1988, the Veterans Administration (VA) is planning to begin production in the vicinity of the BGL and BLM wells. The fact that the VA is planning to reinject their production is encouraging. The reinjection should ameliorate the drawdowns induced by their pumpage. Both the pumping well and the reinjection well should be monitored for production, temperature, and water level. Because the VA production and reinjection wells are close to each other, and to the BLM and BGL wells, those data are particularly important. A study and analyses of the VA data, BGL production and water-level data and water-level data from the BLM well would give a relatively closely controlled network to estimate the local effects of reinjection upon the aquifer's pressure and temperature. The study could yield a model which would offer guidance for development and reinjection elsewhere in the system. Although the hydraulic interconnection in Capitol Mall Wells 1 and 2 are less well understood, a similar analysis of data from those wells should also be undertaken.

Finally, it is also recommended that the reinjection of BGL production be given serious consideration. At this juncture the extent of the effects of reinjection upon temperatures and pressures within the aquifer are not understood; however, reinjection of such a significant amount of geothermal water would be important in helping to maintain pressure within the system.

Additional aquifer testing would be useful in adding to our knowledge of the aquifer characteristics, but considering the present distribution of potential pumping and observation wells, it is not likely to reveal whether or not interconnection exists between the BGL-BLM-CM and BWSWD portions of the aquifer. Because of the rather large distance between the two well groups, BWSWD and BGL-BLM, a prolonged pumping period would be required to create a drawdown at such a distance from the pumping wells. At their present stage of development, both portions of the aquifer are continually in phases of drawdown or recovery, and local withdrawals dominate water-level fluctuations in nearby observation wells.

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We are especially grateful to Mr. Robert Griffiths, Boise Warm Springs Water District Engineer for supplying current and past data on the Warm Springs system; to Mr. Charles Mickelson, City Engineer, and Mr. Robbin Finch for providing production figures for the City of Boise Geothermal System; to Paul Philbrook, Park Designer, Boise Park System for electrical power consumption data, background information on the Quarry View Well and access to the well for monitoring purposes. We wish to thank Mr. W. L. Burnham for sharing some of his broad experience with the geothermal system and for his many thought provoking questions. We appreciate the guidance, help and support from L.K. Mink in developing our proposal for this study and for his helpful suggestions for improving the manuscript. We are indebted to Dr. David W. Small, Boise State University, for developing our computer programs and guiding us through our changing computer needs. We are no less indebted to our students who helped in the data gathering and analysis. Especially, Anna Baumhoff, who donated much of her time to correcting frustrating computer problems, and to Garret Brown and Drew Clemens who were dependable and resourceful in solving monitoring problems which arose frequently. Our thanks also go to Sandra Schmitt who typed the manuscript and cheerfully tolerated the many changes. We are also grateful to Ms. Leah Street, State of Idaho Water Resources, for encouraging this study, handling the budgetary and contractual details, and for her suggestions for improving the study and final manuscript, and for her patience. Finally we wish to express our gratitude to the Department of Energy which funded this study through the Idaho Department of Water Resources.

APPENDIX A

Estimation of Production from Boise City Parks Quarry View Well

Monthly Billing	Kilowatt Hours	Monthly Billing	Kilowatt Hours
1985		**1986**	
April	259		
May	1145	May	849
June	5574	June	6068
July	7312	July	6078
August	4139	August	4990
September	4232	September	2042
October	93		
November	4		
TOTAL 1985	**22,758**	**TOTAL AS OF SEPT BILLING**	**20,027**

To estimate pumpage from the electrical consumption, the following formula was adopted

$$Q = \frac{kwh}{1.8\ (h+p)}$$ where

Q = Acre feet of water pumped

h = depth of water level during pumping

p = pressure head at well in feet of water

kwh = kilowatt hour power consumption

(Young, H.W., Lewis, R.E., an Backsen, R.L., Thermal ground-water discharge and associated convective heat flux, Bruneau-Grand View area, Southwest Idaho: U.S. Geological Survey, Water-Resources Investigation)

For estimation of production a Quarry View a pressure head at the well of 55 psi is used and assumed constant. Fifty-five psi is the designed pressure head for the sprinkler system according to information furnished by P.H. Philbrook, Boise City

Parks. An average depth to water level during pumping of 135 feet was used in the estimates. That pumping level is based upon water level measurements made during this study and drawdowns experienced during pump test in February of 1984 performed by Anderson and Kelly, Inc. For 1985

$$Q = \frac{Kwh}{1.8\ (h+p)} \quad \text{where}$$

$$Q = \frac{22,758}{1.8\ (135 + 126)} \quad Q = 47.89 \text{ acre feet}$$

Multiplying by 43,560 feet/acre and 7.48 gal/ft^3 yields 15.8 x 10^6 gallons produced in 1985

A similar calculation using the 1986 power consumption yields a pumpage estimate of 13.9 x 10^6 gallons for the period late April through the September 1986 billing period.

APPENDIX B

Estimated costs of recommended aquifer studies:

To continue to monitor water-levels in the system using the presently established network and extend the current data base into a period during which production from the aquifer is expected to increase approximately 30 percent.

Travel, salary and operating budget..............$8,000

To expand the monitoring network to include an observation well in the same fracture zone as the City of Boise's (BGL) well field.

Add one pressure-transducer plus a single channel data-logger for a BGL or the BHW Well..............................$2,000

To expand the monitoring network to include an observation well that will monitor pressure levels near the injection well of the V.A. and monitor the Harris and Edwards wells.

Add a second pressure-transducer and single channel data-logger.......................$2,000

Travel, salary, and operating budget.............$3,500

TOTAL $5,500

Study of Fluoride and lithium content of cold water aquifer to determine locations and amount of leakage from geothermal aquifer into cold H_2O.

Fluoride and Lithium Study......................$6,000

Gather precipitation, snow pack data, and stream flow data in potential recharge areas to determine if a correlation of those factors and the geothermal levels exists on a long-term basis

........................$4,000

Inorganic Chemical Analysis and Age dating (Carbon 14) of Harris Well

........................$1,000

Capitol Mall and VA Wells study of production and reinjection temperatures

........................$4,000

Predictive model for equilibrium

Predicated on funding of monitoring network
Review and analysis of all aquifer test data and develop predictive model for equilibrium at various production levels.....................................$10,000

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REVIEW OF GEOCHEMICAL DATA ON

THE BOISE GEOTHERMAL SYSTEM, IDAHO

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CHAPTER 3

REVIEW OF GEOCHEMICAL DATA ON THE BOISE GEOTHERMAL WATER

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Introduction

The primary motivation to examine published chemical analyses of the Boise geothermal waters is to search for parameters that give useful information on the circulation path and the flux of water through the system. Radiocarbon-derived residence times and stable-hydrogen isotope ratios determined by Mayo and others (1984) have been the most informative geochemical data published on these waters. The radiocarbon data show the water has had a relatively long subsurface residence time (4000 - 17,000 years). The stable-hydrogen isotope data are similar to that from other geothermal waters of southern Idaho which Young and Lewis (1982) have interpreted as indicating that these waters originally fell as precipitation during a climate different than the present climate of southern Idaho. The radiocarbon and hydrogen isotope data are consistent in indicating antiquity of the waters.

Many analyses have been published on the chemistry of geothermal waters of southern Idaho. These earlier studies were concerned largely with interpretation of the hydrochemistry in terms of various schemes for estimating "reservoir temperatures" of the waters prior to mixing or cooling during their migration to the surface. Much of this interpretation was done in a "cookbook" manner, using geochemical geothermometer schemes without regard to other factors that could affect the water chemistry. The analyses are a useful data set, but the earlier interpretations are probably of little use, because variables other than temperature also affect the water chemistry.

It seemed worthwhile in this review to investigate the comparative hydrochemistry between several systems and try to explain differences in ion content of the various waters. In order to remove temperature as a major variable in these comparisons, data were selected from waters in the temperature range of 50 to 90^{o}C. This includes the hottest known spring and well waters of southwest Idaho. Locations of springs and wells for which water chemistry is considered in this review are shown in Fig. 3-1.

This review shows that significant differences in the hydrochemistry of waters may be due to aquifer lithology, or the history of encounter of lithologies along the groundwater circulation path, or to variations in the residence times and kinetics of rock-water interaction. Temperature is also a variable, in producing differences, even in the limited range examined, but temperature does not explain many of the differences in chemistry. For example, the Boise geothermal water is distinctive in pH and anion content from waters of most other systems in southern Idaho (Fig. 3-2). Through graphical plotting of various combinations of ions it is possible to separate many of the systems by their characteristic chemistry.

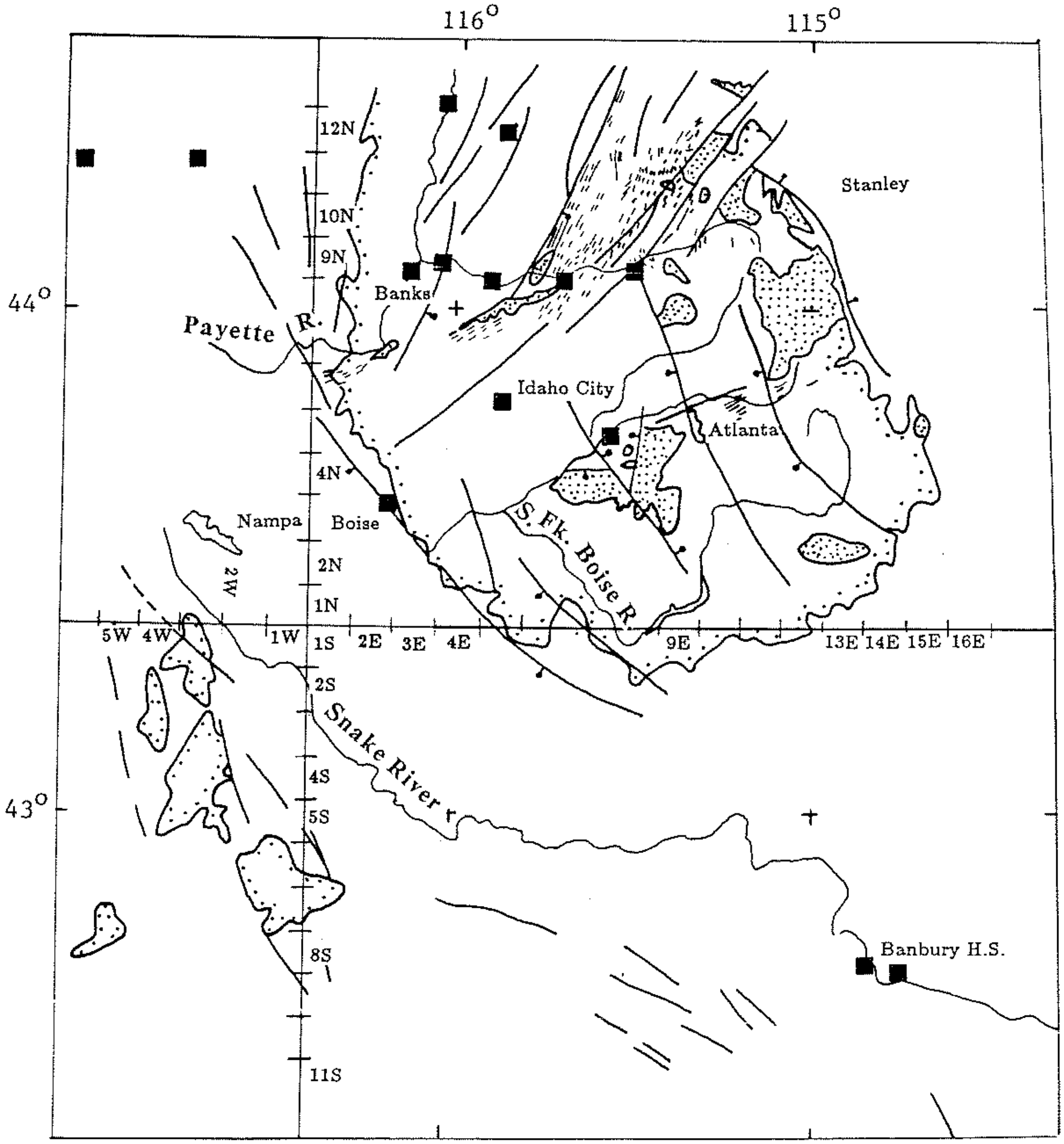

FIGURE 3-1. Map of southwest Idaho showing area of the Cretaceous Idaho batholith (line and dot outline), the Eocene batholiths and stocks (infilled dotted areas) and dike swarms (thin line segments), late Cenozoic major faults with ball and bar on downthrown side (map adapted from Bennett and others, 1985), and locations of geothermal springs and wells discussed in this report.

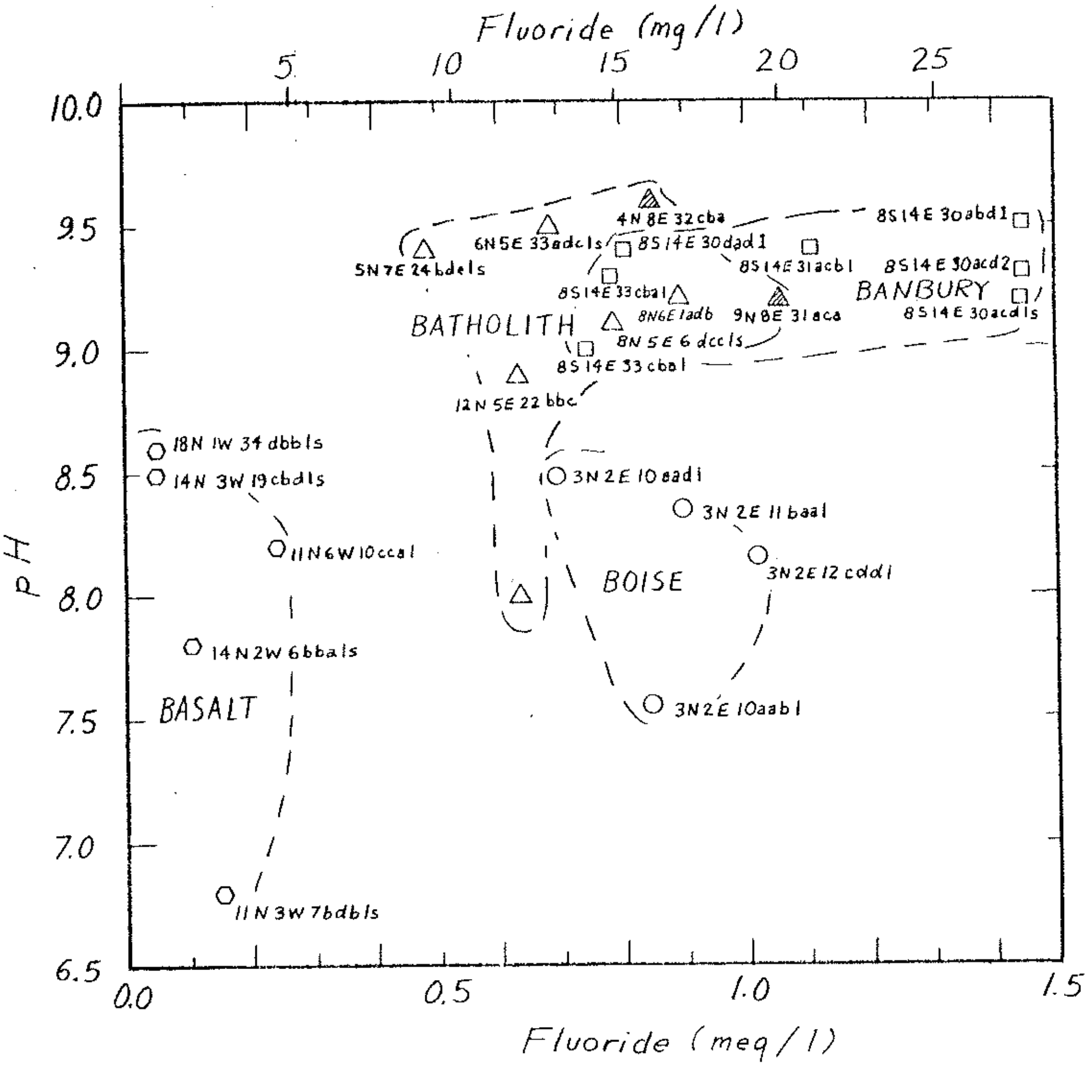

FIGURE 3-2. Plot of fluoride-ion concentration vs. pH of geothermal waters in the Boise area, Banbury area, Idaho batholith region, and selected basalt aquifer warm springs in southern Idaho. Sources of data is in Table 3-1.

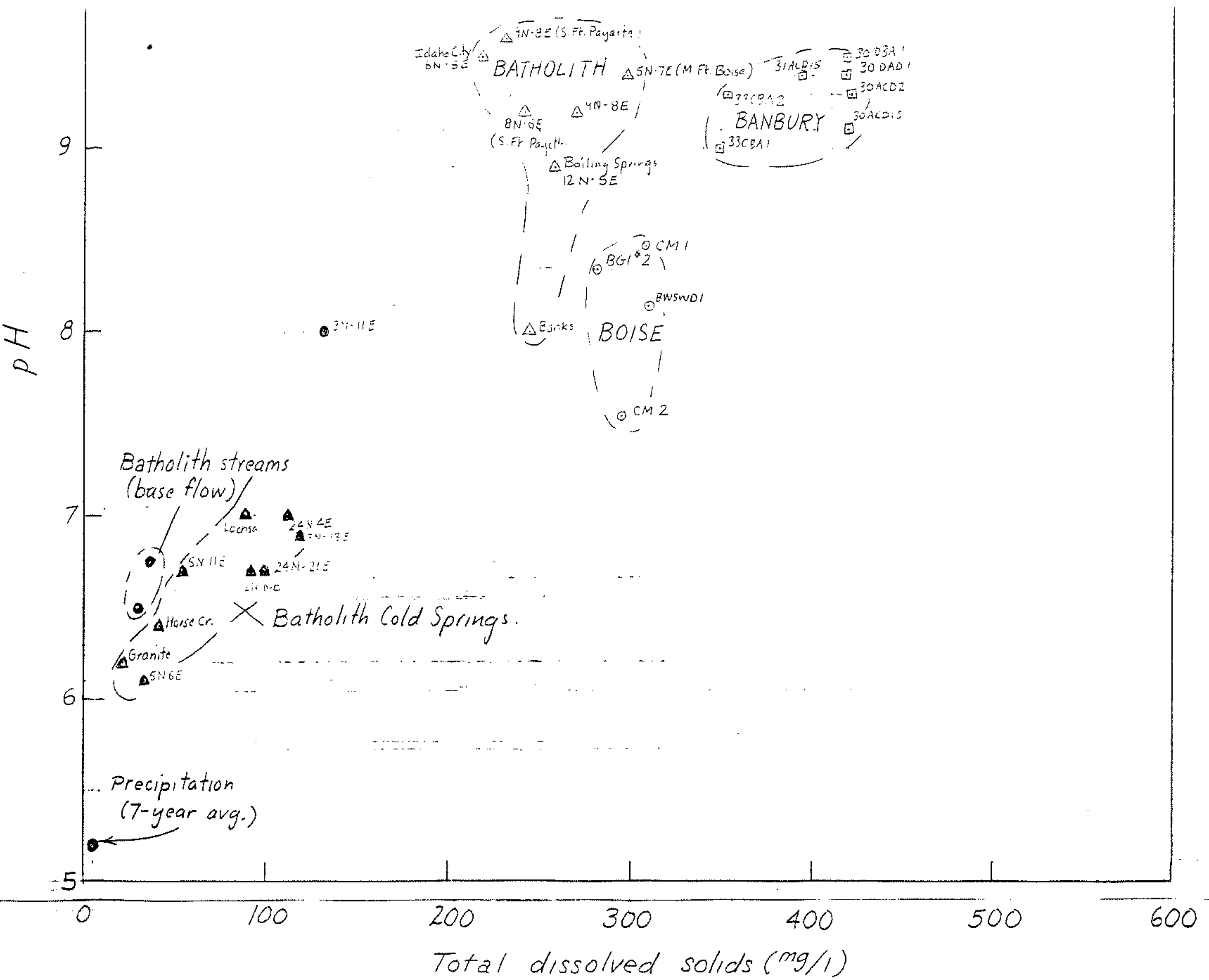

FIGURE 3-3. Plot of total dissolved solids vs. pH for selected geothermal waters in southern Idaho. Data is from Table 3-1 and selected data from Young (1985).

The approach is empirical and mainly for purposes of discovering interesting chemical distinctions between waters from various systems. In future research it may be possible to more precisely interpret their meaning through principles and calculations of mineral-water equilibrium chemistry.

It was hoped that the chemistry would clearly distinguish waters of rhyolite aquifers from waters of the Idaho batholith. They are remarkably similar in major chemistry except for the Boise waters which are of lower pH (Figs 3-2, 3-3). The pH seems to be an indicator of the extent of rock water interaction as shown by a plot of pH vs. total dissolved solids (Fig. 3-3). The pH generally increases with total dissolved solids for springs and wells in the batholith region, a region presumably free of evaporation effects. While no detailed explanation of pH differences is offered in this review, there is a need in future research to geochemically model the water at various temperatures using water-mineral equilibrium calculation programs such as WATEQ (Truesdell and Jones, 1974) now available for personal computers (Rollins, 1987).

Another aspect of the hydrochemistry of systems of similar temperature to the Boise system is the importance of volcanic glass, clay and zeolite minerals in controlling part of the chemistry. As an aid to future studies on the system, several observation of occurrences of these minerals in the aquifer rocks are documented in this paper.

Chemistry of the Boise Geothermal Water

The Boise geothermal groundwater, although relatively low in dissolved solids (TDS range = 240-300 mg/l, see Fig. 3-3, and Table 3-1), has acquired considerable dissolved mineral content since originally falling as precipitation in the region. The water is clearly a $Na\text{-}HCO_3$ type of groundwater as these ions constitute about 70 weight percent of the dissolved mineral matter (Fig. 3-4). By tradition the ionic chemistry of natural waters is plotted and classified on trilinear diagrams of Piper (1944) (See Hem, 1985). The Boise geothermal waters plot into the bicarbonate-anion type and the sodium-cation type fields. In interpreting such diagrams, one must realize that these warm waters cannot simply be compared to plots of cold groundwater chemistry. All geothermal waters are strongly depleted in calcium, magnesium, and potassium relative to ephemeral cold-spring water from similar granite or rhyolite terrain. This presumably occurs because of temperature dependence of exchange reactions with feldspar and clay minerals.

Empirical relationships of $^{(Na)}/_{(K)}$ and $(Ca)^{1/2}/_{(Na)}$ ratios to water temperature form the basis of the Na-K-Ca geothermometer methods (Fournier and Truesdell, 1973) which has been widely used and abused in the literature on low-temperature geothermal waters. Too often this geothermometer is used without regard to the aquifer system lithology, availability of minerals, time availability to achieve equilibrium, and the possibility of dilution and intermixing of waters.

Examination of the chemical analyses of geothermal waters in

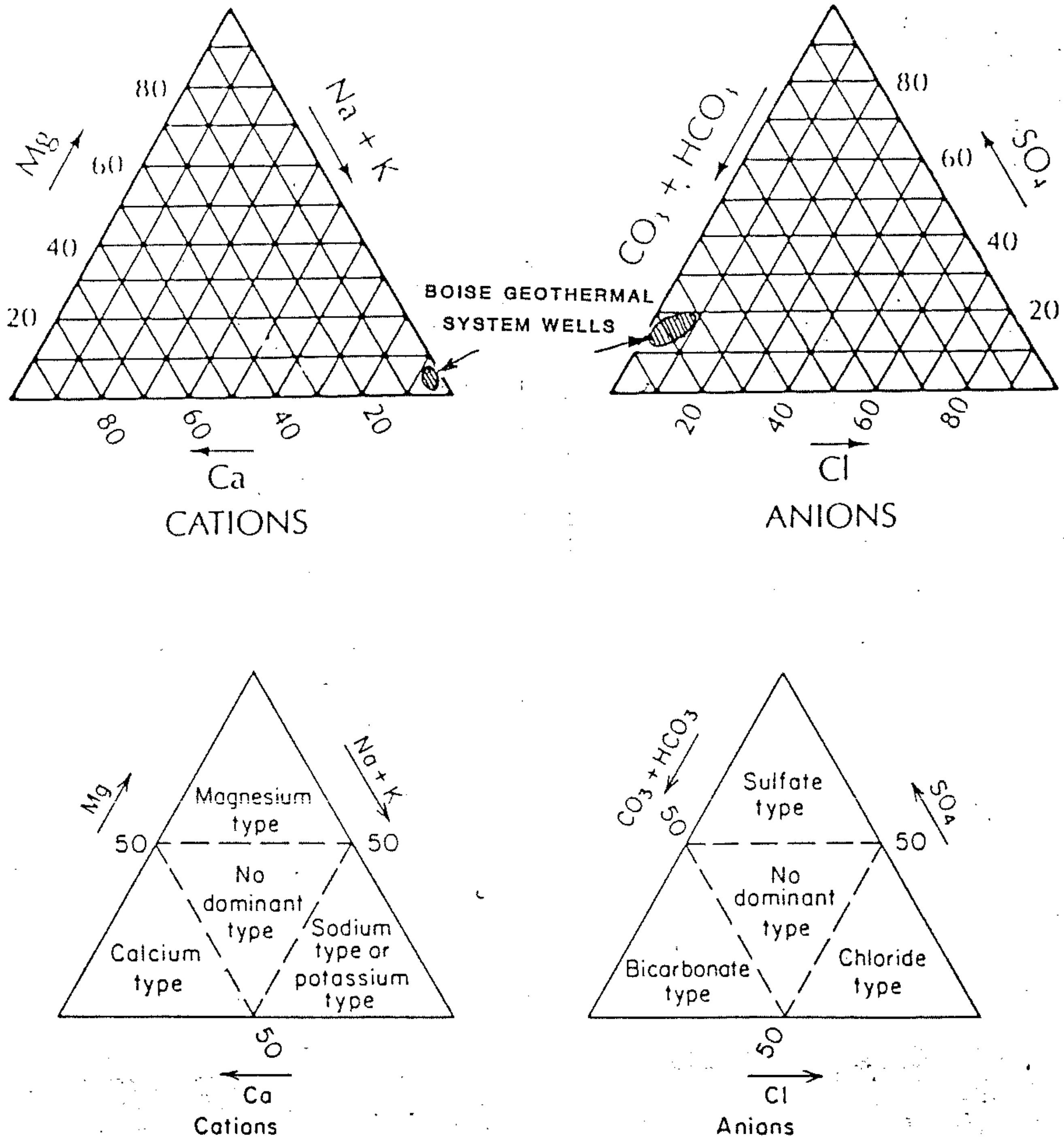

FIGURE 3-4. Piper-ternary plot of major ions in waters of the Boise geothermal area.

TABLE 1. CHEMICAL ANALYSIS OF GEOTHERMAL WATERS FROM WELLS IN THE BOISE ARE AND SELECTED GEOTHERMAL WATERS FROM OTHER AREAS OF SOUTHERN IDAHO
(dissolved chemical constituents expressed in mg/l and in parenthesis as milliequivalents/liter)

BOISE GEOTHERMAL SYSTEM

Spring or Well Name	Sample Date	Flow Rate or Typical Pumpage	Depth of Productive Interval	Temp^{o}C	pH	Ca	Mg	Na	K	Fe	Mn
Boise Warm Springs Water District 3N 2E 12cdd1	8/10/81	P(100-750gpm) F Summer Only	397 (121)	79.5	8.15	3.2 (0.16)	0.10 (0.01)	82.8 (3.60)	1.41 (0.04)		
Boise Geo-thermal Limited Well No. 2 3N 2E 11baa1	8/12/81	F	885 (270)	74.0	8.35	1.6 (0.08)	0.10 (0.01)	86.7 3.77	1.29 (0.03)		
Capitol Mall Well No. 1 3N 2E 10aad1	9/19/81	Flowing? (used a injection)	2,148 (655)	65.0	8.47	1.6 (0.08)	0.10 (0.01)	89.0 3.87	0.70 (0.02)		
Capitol Mall Well No. 2 3N 2E 10aab1		Flowing 250-800 gpm	3,031 924	70.5	7.55	1.3 (0.06)	0.10 (0.01)	79.3 (3.45)	0.59 (0.02)		

BOISE GEOTHERMAL SYSTEM (Cont'd)

Spring or Well Name	Sample Date	HCO_3	CO_3	Cl	F	SO_4	SiO_2	As	B	Li	Hg	Cond.	TDS	Source of Data
Boise Warm Springs Water District 3N 2E 12cdd1	8/10/81	109.8		6.03 (0.19)	19.19 (1.01)	21.6 (0.45)	63.5							Mayo & Others (1984)
Boise Geo-thermal Limited Well No. 2 3N 2E 11baa1	8/12/81	111.1 1.82		7.80 (0.22)	16.91 0.89	25.9 0.54	31.7							"
Capitol Mall Well No. 1 3N 2E 10aad1	9/19/81	126.3 (2.07)		0.89 (0.03)	18.24 (0.69)	16.3 (0.34)	56.5							"
Capitol Mall Well No. 2 3N 2E 10aab1		114.1 (1.87)		10.64 (0.23)	15.96 (0.84)	19.2 (0.34)	55.6							"

WARM SPRINGS IN THE IDAHO BATHOLITH

Spring or Well Name	Sample Date	Flow Rate or Typical Pumpage	Depth of Productive Interval	Temp°C	pH	Ca	Mg	Na	K	Fe	Mn
9N 8E 31aca1s				63.5	9.2	1.4 (0.07)	<0.1	79 (3.40)	1.8 (0.046)		
South Fork (77) Payette		20									
9N 8E 32cba1s	3/27/79	350		64	9.6	1.9 (0.095)	<0.1	70 (3.045)	1.3 (0.03)		
Middle Fork Boise River 5N 7E 24bdd1s	8/3/81			76.0	9.4	1.6 0.080		64 (2.78)	1.7 (0.04)		
South Fork Payette River near 8N 5E 6dcc1s				51.0	9.1	2.9 (0.145)	0.10 (0.01)	77 (3.35)	1.2 (0.05)		
South Fork Payette River near Banks				79.5	8.0	6.1 (0.305)	0.10 (0.01)	120 (5.22)	5.3 (0.23)		
Idaho City 6N 5E 33ooc1s	5/11/81			42.0	9.5	2.0 (0.10)	0.1 (0.008)	65 (2.83)	0.8 (0.02)		
South Fork Payette River 8N 6E 1adb1s	2/27/79			59.5	9.2	2.0 (0.10)	<0.1	75 (3.26)	1.1 (0.028)		
North Fork Payette 12N 5E 22bbc1s	6/5/79			86	8.9	1.4 (0.07)	0.1 (0.008)	74 (3.22)	1.9 (1.56)		
*Boiling Springs	1963			88.0	9.2	2.2 (0.11)	0.0	74 (3.22)	1.9 (0.049)	0.4	0.004

WARM SPRINGS IN THE IDAHO BATHOLITH (Cont'd)

Spring or Well Name	Sample Date	HCO_3	CO_3	Cl	F	SO_4	SiO_2	As	B	Li	Hg	Cond.	TDS	Source of Data
9N 8E 31aca1s		40 (0.656)	31 (1.032)	5.7 (0.161)	20 (1.053)	39 (0.812)	83.0	0.002	0.020	0.120	$<10^5$	369	270	Young (1985)
South Fork (77) Payette														
9N 8E 32cba1s	3/27/79	29 (0.47)	29 (0.81)	3.9 (0.11)	16 (0.84)	39 (0.81)	67	0.002	0.050	0.110	$<10^5$	321	232	"
Middle Fork Boise River 5N 7E 24bdd1s		16 0.26	46 1.53	2.4 0.67	9.2 (0.48)	28 (0.58)	100 1.66	0.001	0.04	0.09	0.0002	297	261	Y "
South Fork Payette River near 8N 5E 6dcc1s		67 (1.10)	20 (0.67)	7.8 (0.21)	15 (0.78)	36 (0.74)	63 (1.05)	0.006	0.11	0.13	$<10^5$	343	245	Y "
South Fork Payette River near Banks		160 (2.62)	0.0	39 (1.09)	12 (0.63)	12 (0.25)								Y "
Idaho City 6N 5E 33ooc1s	5/11/81	41 (0.672)	36 (1.20)	2.7	13	22	57	.035	.050	.030	0.1×10^7	317	219	"
South Fork Payette River 8N 6E 1adb1s	2/27/79	39 (0.67)	29 (1.20)	6.7 (0.1)	17 (0.89)	38 (0.79)	63 (1.05)	0.006	0.080	0.100	<0.01	343	242	"
North Fork Payette 12N 5E 22bbc1s	6/5/79	95 (1.56)	18 (0.60)	11 (0.31)	12 (0.63)	12 (0.25)	86 (1.43)	0.01	0.09	0.08	0.0002	331	259	"
*Boiling Springs	1963	106 (1.74)		14 (0.39)		12 (0.25)	81 (1.35)	<0.1	0.1					White 1967

BANBURY HOT SPRINGS AREA

Spring or Well Name	Sample Date	Flow Rate or Typical Pumpage	Depth of Productive Interval	Temp°C	pH	Ca	Mg	Na	K	Fe	Mn
Banbury H.S. 8S 14E 30acd1s	2/8/78	40		70.5	9.1	1.2	<0.1	140	1.1		
Banbury HS 8S 14E 3acd2	8/15/79	120	420	72.0	9.3	0.9	0.1	140	1.2		
Banbury HS 8S 14E 30dad1	3/26/79	9	700	62.0	9.4	0.7	<0.1	150	1.4		
Banbury HS 8S 14E 30dba1	4/25/79	650	450	71.5	9.5	1.5	0.1	140	1.4		
Banbury HS 8S 14E 31acb1S	4/5/79	290	290	57.0	9.4	0.9	<0.1	130	1.5		
Banbury HS 8S 14E 33cba1	2/8/78	60	110	59.0	9.0	1.1	<0.1	100	1.5		
Banbury HC 8S 14E 33cba2	4/4/79		342	59.0	9.3	1.1	<0.1	110	1.6		

BANBURY HOT SPRINGS (Cont'd)

Spring or Well Name	Sample Date	HCO_3	CO_3	Alkalinity or $CaCO_3$	Cl	F	SO_4	SiO_2	As	B	Li	Hg	Cond.	Source of Data
Banbury HS 8S 14E 30acd1s	2/8/78	70	45	132	50	27	32	89	62	440	60	<0.1	601	Lewis and Young (1982)
8S 14E 3acd2	8/15/79	59	52	135	51	27	35	86	52	470	60	0.1	633	"
8S 14E 30dad1	3/26/79	56	55	138	48	15	35	84	43	490	50	<0.1	634	
8S 14E 30dba1	4/25/79	56	55	138	51	27	33	82	42	510	60	0.2	646	
8S 14E 31acb1S	4/5/79	59	58	145	34	21	34	86	48	340	40	0.1	566	
8S 14E 33cba1	2/8/78	90	35	132	25	14	28	100	40	230	40	<0.1	454	"
8S 14E 33cba2	4/4/79	78	46	141	23	15	27	88	36	260	40	0.1	466	"

HOT SPRINGS ARKANSAS

Spring or Well Name	Sample Date	Flow Rate or Typical Pumpage	Depth of Productive Interval	Temp°C	pH	Ca	Mg	Na	K	Fe	Mn
Hot Springs Arkansas Spring No. 49				61.8	6.95	44 (2.20)	48 (0.39)	3.1 (0.11)	1.5 (0.04)		

HOT SPRINGS ARKANSAS (Cont'd)

Spring or Well Name	Sample Date	HCO_3	CO_3	Cl	F	SO_4	SiO_2	As	B	Li	Hg	Cond.	TDS	Source of Data
Hot Springs Arkansas Spring No. 49		155 (2.54)		1.9 (0.05)	0.2 (0.01)		8.2 (0.17)				184	199	0.00044	$10^{13.36}$

WARM SPRINGS & WELLS IN BASALT AQUIFERS, WESTERN IDAHO

Spring or Well Name	Sample Date	Flow Rate or Typical Pumpage	Depth of Productive Interval	Temp^{o}C	pH	Ca	Mg	Na	K	Fe	Mn
16M 6W 10cca1 Flowing Well H2S Odor Miocene Basalt	6/28/72	0.3	400	70	8.2	2.7 (0.13)	0.0	160 (6.96)	5.1 (0.13)		
11M 3W 7bd1s Two Springs Miocene From Basalt	6/30/72	10	1	87	6.8	27 (1.34)	0.7 (.058)	300 (13.05)	19 (0.48)		
14N 3W 19cbd1s	6/27/72	58		50	8.5	8 (0.40)	0.8 (.066)	80 (3.48)	1.9 (0.049)		
14N 2W 6bda1s Springs from Miocene Basalt hydrogen sulfide odor	6/28/72	431		70	7.8	17 (0.85)	0.1 (.008)	200 (8.70)	3.8 (0.097)		
Starkey Hot Springs (Adams Co) 18N 1W 34dbb1s Seven Spring Vents Basalt?		492		55	8.6	4.5 (0.22)	0.0	86 (3.74)	1.6 (0.041)		

WARM SPRINGS & WELLS IN BASALT AQUIFERS, WESTERN IDAHO (Cont'd)

Spring or Well Name	Sample Date	HCO_3	CO_3	Alk-as $CaCO_3$	Cl	F	SO_4	SiO_2	As	B	Cond.	TDS	Source of Data
16N 6W 10cca1 Flowing Well H2S Odor Miocene Basalt	6/28/72	92 (1.507)	19 (0.63)	107	55 (1.55)	4.6 (0.24)	150 (3.123)	170 (2.83)			698	612	Young & Mitchell (1973)
11N 3W 7bdb1s Two Springs Miocene From Basalt	6/30/72	198 (3.24)		162	190 (10.00)	2.9 (0.15)	270 (5.62)	170 (2.83)			1480	1080	"
14N 3W 19cbd1s	6/27/72	81 (1.33)	1 (0.03)	68	05 (0.79)	0.8 (0.042)	110 (2.29)	55 (0.915)			406	314	"
14N 2W 6bda1s Springs From Miocene Basalt hydrogen sulfide odor	6/28/72	24 (0.39)	20 (0.67)	53	140 (7.37)	1.9 (0.10)	200 (4.164)	72 (1.20)			1000	667	"
Starkey Hot Springs (Adams Co) 18N 1W 34dbb1s Seven Spring Vents Basalt?		60 (0.98)	6.0 (0.20)	59	14 (0.39)	0.90 (0.047)	150 (1.17)	56 (0.93)			502	348	"

IDAHO BATHOLITH REGION PRECIPITATION AND STREAM FLOW

Spring or Well Name	Sample Date	Flow Rate or Typical Pumpage	Depth of Productive Interval	Temp°C	pH	Ca	Mg	Na	K	Fe	Mn
Precipitation in the Silver Creek Study Area of the Idaho Batholith (7-year weighted average 65% snowfall, 35% rain, 78-84 - from J. Clayton, USFS					5.2	0.62	0.05	0.10	0.17		
Streams - typical base flow value in the Silver Creek Study Area											
Stream 1					6.7	8.00	0.732	3.65			
Stream 2					6.75	6.37	0.775	5.44			
Stream 3						1.40	8.74	0.98			

IDAHO BATHOLITH REGION PRECIPITATION AND STREAM FLOW (Cont'd)

Spring or Well Name	Sample Date	HCO_3	CO_3	Cl	F	SO_4	SiO_2	As	B	Li	Hg	Cond.	TDS	Source of Data
Precipitation in the Silver Creek Study Area of the Idaho Batholith (7-year weighted average 65% snowfall, 35% rain, 78-84 - from J. Clayton, USFS				0.21		1.84	nd <<1ppm							J. Clayton USFS, Boise
Streams - typical base flow value in the Silver Creek Study Area														
Stream 1				0.64		2.77	20.94							"
Stream 2				0:88		4.6	28.2							
Stream 3				1.06		2.94	27.0							

the range of 50^{o} to 90^{o}C shows that the major differences among waters from different areas are mostly in their anion content, pH, and potassium ion. Fluoride vs. pH (Fig. 3-2) most clearly separates waters from various areas and inferred lithologies into fields with the least amount of overlap. Other plots are shown in Figures 3-5 through 3-7.

Many of the geothermal waters of southern Idaho have circulated through volcanic rocks. Perhaps the most reactive and significant component of these rocks is volcanic glass of several compositions. Previous studies have examined evolution of waters with respect to discrete silicate minerals but have neglected consideration of natural glass. Chemistry of the water of the Boise system, Bruneau-Grandview, and Banbury area, however, may have been determined largely by interaction of water with rhyolite glass.

Fluoride content

It is well known that many geothermal waters are high in fluoride ion; many exceed the safe limit (1.0 - 4.0 mg/l) for drinking water supply. Nordstrom and Jenne (1977) show that fluorite solubility provides equilibrium control on dissolved fluoride activity in geothermal water. They maintain that waters unsaturated with respect to fluorite are those that have undergone dilution by non-thermal waters, as shown by correlation of decreased conductivity and temperature with unsaturated fluorite activity product for the natural waters. They do not, however consider variations in aquifer lithology of the 351

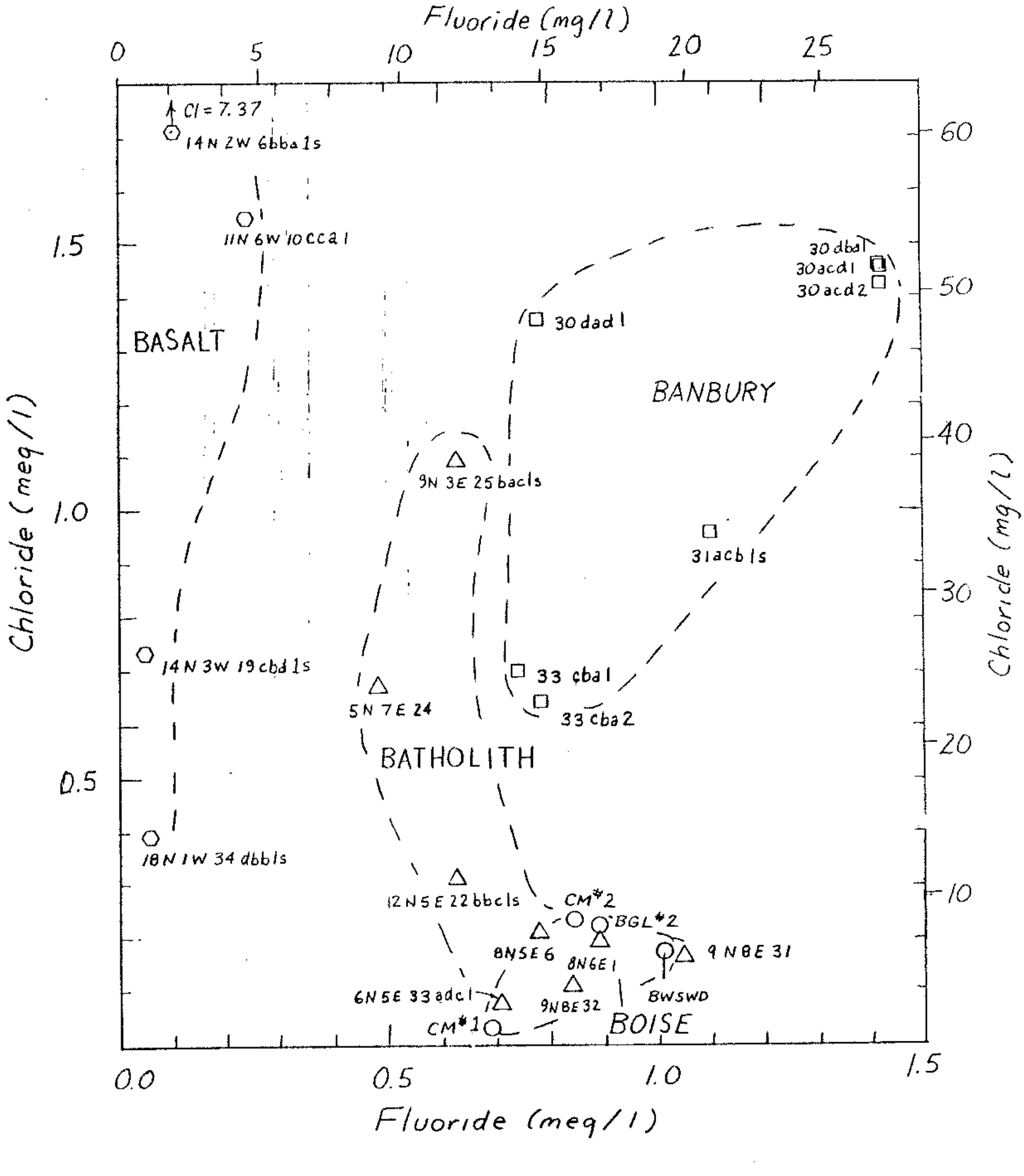

FIGURE 3-5. Plot of fluoride-ion concentration vs. chloride-ion concentration for selected geothermal waters in southern Idaho.

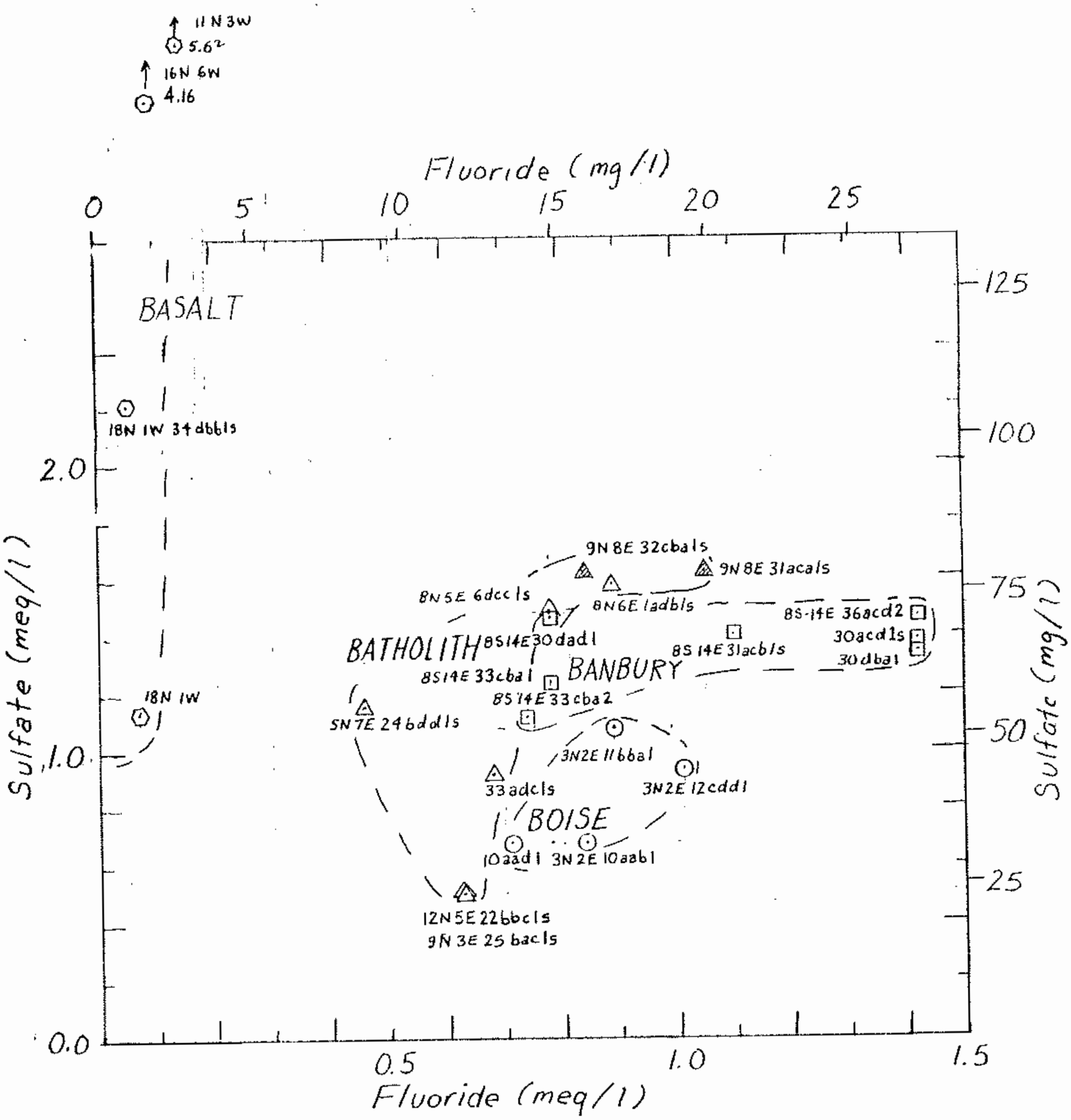

FIGURE 3-6. Plot of fluoride-ion concentration vs. sulfate-ion concentration for selected geothermal waters in southern Idaho.

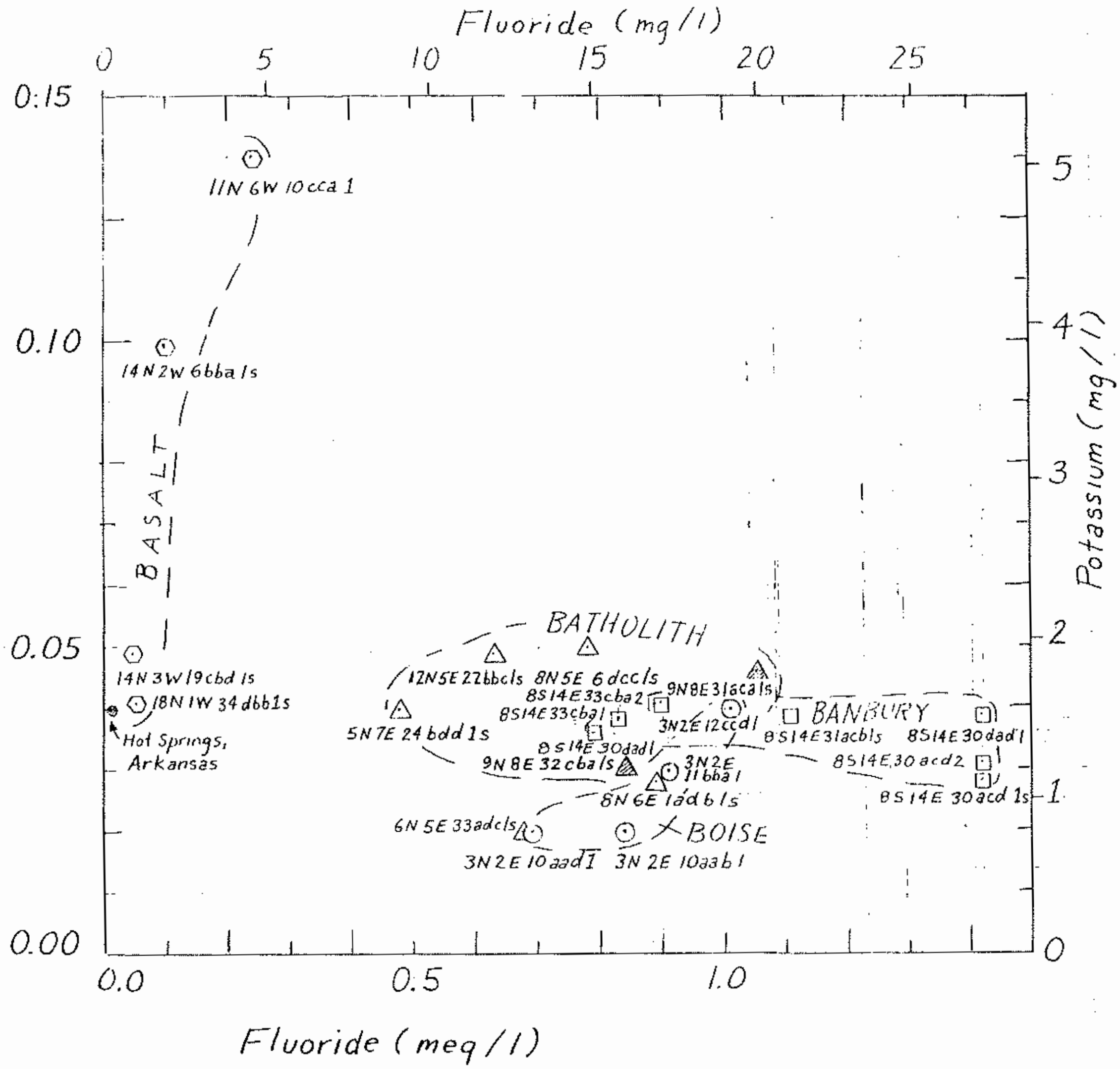

FIGURE 3-7. Plot of fluoride-ion concentration vs. potassium ion-concentration for selected geothermal waters in southern Idaho.

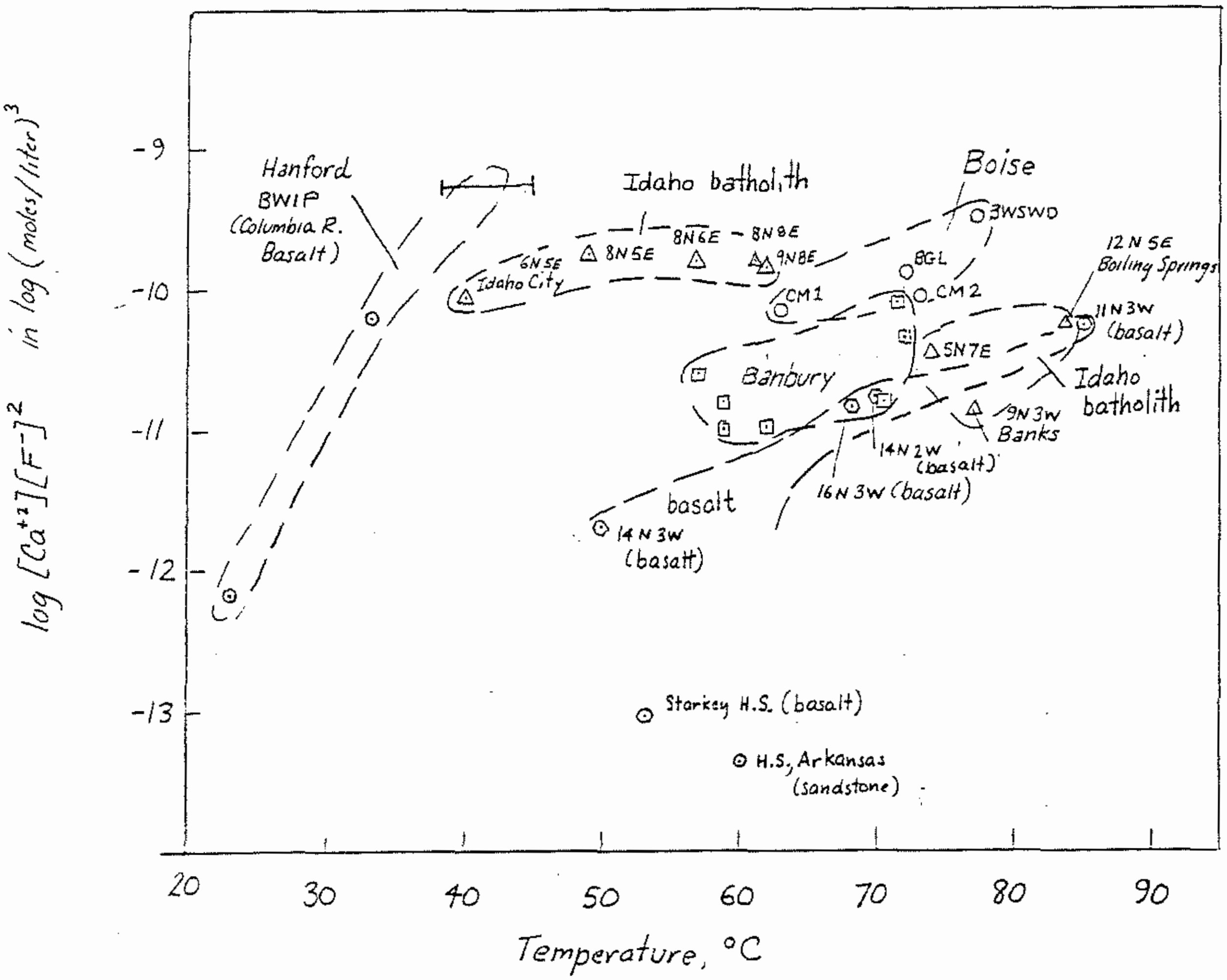

FIGURE 3-8. Plot of concentration product with respect to fluorite solubility vs. temperature for selected geothermal waters. Note that the product does not exceed $10^{-9.8}$ suggesting that this is the concentration product for saturation with respect to fluorite (CaF_2). Note that this is not the thermodynamic activity product, which must be calculated by taking into account the ionic strength of waters. However, ionic strengths are relatively low, and such a calculation should not significantly change values shown here. Data from table 3-1 of this review except data from Columbia River Basalt which is from an article authored by D.E. White referenced as Waste Isolation Systems Panel of the National Research Council (1983).

reported analysis they studied. In order to test the effect of lithology, the concentration product $(Ca^{+2})(F^{-})^{2}$, is plotted against temperature for waters of southern Idaho (Fig. 3-8). The highest values of concentration products are about $10^{-9.8}$ (moles/liter)^{3}, which probably indicates saturation with respect to fluorite for the Boise geothermal waters and some Idaho batholith waters.

Because the waters tend to establish equilibrium with respect to fluorite and calcite and most geothermal waters are high in pH, calcium is removed from solution to satisfy calcite equilibria. Removal of calcium causes more fluoride to come into solution to satisfy fluorite equilibra; therefore, the high pH apparently favors high fluoride. The equilibrium relationships are complicated because they involve the many reactions of the carbonate system, as well as fluorite. This subjective discussion is simply a suggestion for later research, on the various factors which may affect the fluoride content, some of which are reviewed in more detail in the following discussion.

Fluoride is a significant component of geothermal waters from granitic rocks of the batholith, and from rhyolite aquifers of Boise and the Banbury Hot Springs area. In these waters flouride is anomolously high and ranges from 8 to 26 mg/l. Fluoride content of waters from basalt aquifers in southern Idaho is notably low. Geothermal spring waters from the Idaho batholith are also high in fluoride (Fig. 3-2), so that fluoride does not clearly distinguish waters that resided only in granitic rocks from waters that resided rhyolites. However, waters from western batholith (west of range 7E) are somewhat lower than waters from the eastern batholith (east of Range 7E) and the rhyolite waters (Fig. 3-1, 3-2).

Silicic volcanic rocks are generally high in fluoride content. Coats and others (1963) show a mean value of 520 ppm fluorine in silicic volcanic rocks of the western U.S., and mean of 970 ppm for 47 samples from what he termed the "Shoshone rhyolite province of Northern Nevada and Southern Idaho" (now regarded informally as the "hot-spot rhyolites"). While fluoride analysis of the specific Idavada group rhyolites that lay in the subsurface area are not available, these rocks are in that same province and it is likely that these rocks are similarly high in fluoride. Fluoride in granitic rocks is quite variable, ranging typically from 500 to 800 ppm (Koritnig, 1972). Unfortunately no analysis from suites of rocks from Idaho or the Sierra Nevada are published. Fluoride would presumably be released to percolating groundwater from coarse crystalline rocks by alteration of biotite. Release from small apatite crystals surrounded by quartz or feldspar might be slow, because of the relative resistance to alteration of these minerals.

Are waters from granitic rocks typically high in fluoride? Feth and others (1964) report values ranging from 2 to 8.5 mg/l for 5 springs in the Sierra Nevada batholith ranging in temperature from 49^{o} to 65^{o}C. These values are lower but more comparable to the western Idaho batholith waters (9-11 mg/l). Waters from the eastern Idaho batholith (east of Range 7E) range from 12 to 20 mg/l. An explanation for the fluoride differences in Idaho geothermal waters was suggested to the author by Earl Bennett (Idaho Geological Survey). The eastern batholith contains, in addition to Cretaceous plutonic rocks, abundant shallow intrusive rocks of the so called "Eocene volcanic-plutonic event" (Armstrong, 1974). Although total fluorine analysis are not available for these rocks, Bennett and Knowles (1985, p.87) analyzed biotite from these intrusive rocks and found that biotite from the Tertiary rocks is 3 to 4 times higher in fluorine content than biotite from the Cretaceous intrusive rocks. The Tertiary biotites they analyzed are typically 4 per cent fluorine. Distribution of the Tertiary intrusive rocks and springs for which analysis are used in this report is shown in Figure 3-1.

The behavior of fluorine in the magmatic environment is beyond the scope of this discussion, but it is quite mobile and appears to move to more silicic magmas in magmatic differentiation processis. It is apparently immobilized by available phosphorous as apatite or by the formation of micas and amphiboles. Noble and others (1967) indicate fluoride is high in unaltered silicic glass, but that about 1/2 of it is lost in devitrified glasses.

The Eocene shallow intrusives of the Idaho gold porphyry belt, the Challis volcanics, and the Eocene "pink granites" are all capable of yielding fluoride to thermal groundwater, and probably account for the geographic increase in fluoride in the eastern group of Idaho batholith springs.

Fluoride content of geothermal water from basalt areas of southern Idaho appear to be quite low in fluoride (Fig. 3-2), and seemingly are distinguishable from waters that have equilibrated with rhyolite. This finding is similar to work of Arnorsson and others (1983) who found that geothermal waters in Iceland percolating through acidic (silicic) volcanic rocks seem to saturate with respect to fluorite equilibria, whereas waters from basaltic rocks in Iceland are undersaturated. Fluoride contents of 70°C waters from acidic rocks are 4 - 12 mg/l, and 0.2 to 1.5 mg/l in waters from basaltic rocks. They believe the differences may be due to greater fluoride content of the acidic volcanic rocks, but suggest that ion exchange equilibria involving OH^- and F^- may control activity of fluoride in waters flowing through basaltic rocks. In the case of acidic volcanic rocks, they suggest the rate of leaching of fluoride may be too rapid for consumption of this fluoride by hydroxide-bearing silicates.

An exception to the conclusion that water from basalts are low in fluoride is found in the Columbia River Basalt section at the Hanford Nuclear Waste Repository site, which shows very high fluorine in the groundwaters below a depth of 2200 ft. At a depth of 2200 - 3100 ft, the temperature reaches 57°C, and fluoride content is 17.5 to 22.9 mg/l (Waste Isolation Systems Panel, 1983, p.169). Fluoride analysis on the specific rocks from the wells have not been released, but data compiled by Koritnig (1981) show fluorine in 5 Iceland basalts averaged 180 ppm, whereas analyses of 16 Columbia River Basalts by Seraphim (1951) averaged 540 ppm, similar to values of rhyolites. Waters from the Columbia River Basalt with high fluoride content also have radiocarbon based residence times in excess of 33,000 years, which is presumably ample time for rock-water interaction, development of high pH, and to reach equilibrium with respect to fluorite.

Although fluoride seems to distinguish waters from rhyolite aquifers in southern Idaho, this may not be a valid generalization. It should be worthwhile in future study to examine a broader group of fluoride analysis in geothermal waters, and also use the conservative nature of fluoride in waters diluted by mixing of geothermal water with cold waters to detect the extent of leakage and mixing of geothermal waters into the cold-water aquifers.

Chloride content

Boise geothermal water and some thermal springs from the batholith are anomolously low in chloride (0.9 to 10.6 ppm) relative to water from the Banbury area and from basalt aquifers (Fig. 3-5). In fact, some chloride in some thermal waters is not much greater than chloride in the base flow of mountain streams (Table 3-1). Chloride content in waters from basalt aquifers, Banbury area and some western batholith warm springs are much higher, from 10.5 to 50 ppm. Significance of the large chloride variation is not at all understood. However, most of the thermal springs from granitic rocks in the vicinity of Boise also have low chloride. This suggests that granitic rocks are the main deep circulation path for the Boise geothermal waters. Certainly, the graph of fluoride vs. chloride (Fig. 3-5) is a means of chemically fingerprinting and distinguishing these southern Idaho geothermal aquifer systems using routine analysis.

Sulfate content

Sulfate ion in the Boise waters is present at levels of 0.34-0.54 meq/l (16-26 mg/l). It is less than most thermal spring waters at equivalent temperatures in the Idaho batholith and the Banbury area (Fig. 3 -6). Sulfate in thermal springs from basalt is clearly much higher than both batholith waters and rhyolite waters (Fig. 3 -6) . Sulfate is typically 3 to 10 times more abundant in basaltic glasses than in rhyolite glasses, whereas chlorine is more variable, but also appears to be more abundant in basaltic glass, Anderson (1974, Table 1). This may explain the sulfate differences in waters derived from the different volcanic aquifers in southwest Idaho.

It is possible that some of the reported sulfate was actually dissolved H_2S that became oxidized during sampling and analysis. There is a slight H_2S odor in the Boise waters, but H_2S gas is so odorous that it can be detected as a gas in air by human smell at levels of 0.002 ppm (USEDA, 1976, p.213). Henry's law computation using $K_H = 6 \times 10^{-2}$ mole/kg bar, implies that 0.002 ppm H_2S in air would be 10^{-4} mg/l in water, at equilibrium. At pH 8.5, such an H_2S content would be in equilibrium with about 0.1 mg/l HS^- and 10^{-7} mg/l S^{-2}. In other words, if one can smell H_2S, one can conclude that the total sulfide species in water is at least 0.1 mg/l and that HS^- ion is dominant.

By way of comparison, Arnorsson and others (1983) report typical sulfate and sulfide in geothermal water of similar temperature as 60-80 mg/l sulfate and sulfide from <0.01 to 1.7 mg/l. These analyses are on waters from basalt aquifers in Iceland. They also point out that careful analyses of sulfate, sulfide, and pH at the well site does allow an estimate of the redox potential of the waters. Such well sight analyses might be employed in further geochemical study of Idaho geothermal waters.

pH of the water

The Boise Geothermal waters have an alkaline pH ranging from 7.5 to 8.5 (Fig. 3-2.), They are distinguishable from most hot springs waters from the Idaho and Sierra Nevada batholith and the Banbury rhyolite aquifer, because they are not as alkaline as those waters. The alkaline pH of thermal water is explained by Feth and others (1964) in this way: "At somewhat elevated temperatures, the chemical agressiveness of the water is increased and the H^+ion content is exhausted by vigorous hydrolosis of silicate minerals thus raising the pH."

Pure water at 25°C has a pH of 7, but pure water of 60°C has a pH of 6.5, or 3.2 times higher in H^+ ion concentration (Matthes, 1982, p.19) so that although H^+ activity increases with temperature, the hydrolysis reactions of silicate minerals are endothermic and consume available H^+ ions, so that the extent of silicate mineral - water interaction is probably reflected in the pH. In Fig. 3-3 it is shown that total dissolved solids in waters circulating through igneous rocks increases with pH. This relationship is probably controlled by the extent of the mineral-water reactions.

The pH is controlled by products of hydrolysis such as HCO_3. Just why the Boise waters are generally lower in pH than batholith or other rhyolite aquifer rocks is not known. Also unusual is the lower pH (7.5) of water from the Capitol Mall no. 2 (CM2) well, the deepest well water of the system. Without repeated well-site measurements of the CM2 water it doesn't seem worthwhile to digress greatly; however, pH is a simple measurement and variations among the system water may be informative. Perhaps residence time in rhyolite is a factor in producing differences in hydrogen ion concentration.

Cation content of the waters

Boise waters are slightly lower in potassium than waters from some batholith springs (Fig. 3-7). This may be an effect of interaction with rhyolite glass as discussed below. There is great similarity in sodium, calcium, and magnesium content of Boise, Banbury, basalt aquifers, and thermal waters from granitic rocks. This suggests that evolution of major cations is controlled by factors other than lithology of igneous aquifer rocks.

Feth and others (1964) have indicated that most thermal waters from granitic rocks they examined are supersaturated with respect to both calcite and dolomite, and that Ca and Mg may be depleted relative to sodium by precipitation of these minerals.

They also suggest that magnesium as well as potassium may be incorporated into clay minerals formed during alteration of igneous rocks. Equilibrium relationships for these water-rock-clay mineral interactions have not been extensively studied in the range of low-temperature geothermal waters, and little more can be inferred from available literature. Waters of this study, however, also have interacted with volcanic glass, and some data is available on rhyolite glass-water interaction, largely because of interest in the geochemistry of waters at the Nevada Nuclear Test Sites. These data however, are mostly at 25°C, and not at 60°C.

White (1979) has examined the interaction between rhyolite glass and 25°C groundwater in tuffaceous rocks of southern Nevada. In the hydration and leaching of glass, Na_2O is quite mobile. The glasses preferentially absorb or retain hydrogen, potassium, and to a lesser extent calcium relative to sodium. On the basis of Truesdell's (1966) experiments, White (1979) concludes that vitric tuffs should produce sodium-rich groundwaters. Noble (1965) and Truesdell (1966) show that glasses interacting with groundwater preferentially retain the larger potassium ion and release the smaller sodium, calcium, and magnesium ions.

Importance of lithium analyses

Lithium concentrations have not been published for the Boise geothermal waters. In order to determine if it might aid in distinguishing waters from various sources, lithium values vs. pH

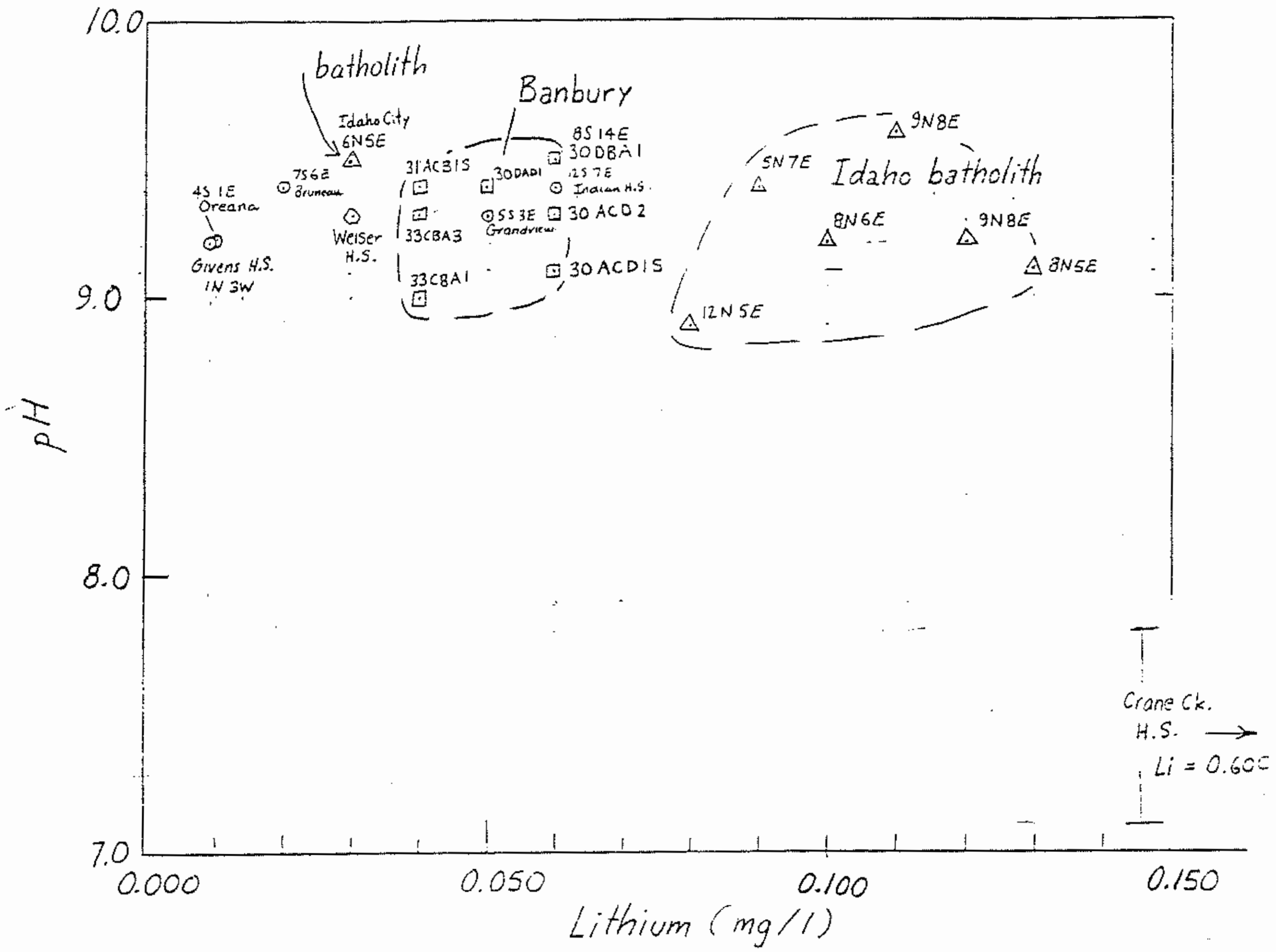

FIGURE 3-9. Plot of lithium concentration vs. pH for selected geothermal waters of southern Idaho.

published by Young and Lewis, 1982; Rightmire and others, 1976; Lewis and Young, 1982; and Young, 1985 are plotted on Figure 3-9. All but one of the batholith waters are distinguishable by lithium content from the waters of rhyolite systems of Banbury Hot Springs, and the Bruneau-Grandview area. Since this is the only distinctive ion of batholith waters, it is imperative that future analysis of the Boise water include lithium. Lithium might serve as an indicator of temperature, residence time, extent of rock water interaction, or of a coarsely-crystalline acid plutonic- rock lithology (i.e. lithium may reside between crystals in a course crystlalline rock, whereas in a glassy rock, it may be within the glass, and not as accessable to solution by water).

High lithium concentrations and low sodium/lithium ratios have been proposed as indicators of high temperature geothermal reservoirs by Foullac and Michard (1981). They suggest that the lithium - temperature relationship is not likely to be related or limited by chemical equilibrium of lithium minerals, lithium minerals are rare in hydrothermal environments. However, Keith and others (1983) found high lithium contents in some smectite clay and clinoptilolite zeolite minerals in hydrothermally altered rhyolite in the Yellowstone area, suggesting that secondary minerals from alteration of glass may take up some lithium. Studies of hydrothermal alteration and fluid-rock interactions of rhyolite in the Valles Caldera by White (1985) indicate that lithium is conservative in fluids, and an indicator of the extent of rock-fluid interaction in the temperature range of 50 to 300°C. For these reasons, it is an important determination in the hydrochemistry of geothermal waters.

Hydrothermal Minerals Associated with the Boise Geothermal System

An interesting mineral assembledge has been observed in cuttings during drilling of geothermal water wells, and in initial water flow from one of the wells. Because the geochemistry of the water is rather well known, it seems worthwhile to document observations on these minerals by Will Burnham and Spencer Wood, who examined many of the drill cuttings and surface outcrops of rocks in uplifted fault blocks that were formerly geothermal aquifers. A thorough study of hydrothermal minerals associated with the system would be a useful countribution to our knowledge of the geochemistry of this type of epithermal system. Cuttings are available and a study using x-ray diffraction and scanning electron microscope techniques is encouraged, but was not attempted by the authors.

Cuttings from the wells contain abundant pyrite or marcasite(?) encrustations, thin red and green jasper veins, white chalcedony veins, and vesicle fillings of chalcedony, calcite and zeolite. Zeolite and clay alteration products presumably replace parts of the glassy rhyolite, but these rock cuttings from the rhyolite aquifer have not been analyzed by x-ray diffraction. Pyrite is first encountered at a depth of 200 ft. in the Kanta and Boise Geothermal Limited wells as encrustations on sand particles in porous beds and as crystals and coatings in fractures in claystones. Pyrite persists throughout the deeper cuttings. Jasper and chalcedony are common in all cuttings from the rhyolite. Zeolites (exact minerals unknown) are in the deeper basalt units in the Kanta well and in the rhyolite of all wells.

The most unusual mineral occurrence happened during the initial production testing at Boise Geothermal Ltd. well No. 2. Accompanying the flow of artesian hot water were abundant thumbnail-size tabular crystals and crystal fragments of clear stilbite. Also produced were pieces of soft white clay having the appearance of "Ivory Soap". Embedded in the white clay were crystals of clear stilbite (Will Burnham and Jack Kelly, Anderson & Kelly Consultants, Boise; personal communication). The very pure appearance of the stilbite and clay-mineral strongly suggests they were precipitated as mineral growths on the walls of fractures. The large flow velocity of the well test apparently scoured these loosely-adhered minerals from fracture walls into the turbulent stream of water in the well bore.

Stilbite, calcite, and possibly heulandite (?) occur in veins about 10cm thick in silica-cemented sandstone on Table Rock (NW 1/4, NE 1/4, Section 19, 3N, 3E) (Burnham and Wood, 1985; p. 17). Unlike the stilbite produced from the wells; however, abundant clay is not associated with these veins.

The clayey material produced from the Boise Geothermal Well No. 2, has a strong 14-A^{o} x-ray diffraction peak identifying it as one of the smectite group of clay minerals (S.H. Wood, unpublished analyses). Most likely, it is a Na-montmorillonite, but a chemical analysis has not been done.

Zeolite in Low-Temperature Hydrothermal Systems

Little work has been done on the geochemistry of relatively low-temperature groundwater systems (<100^{o}C), nevertheless, these systems clearly alter the rock materials and form new minerals in a hydrothermal environment. Boles (1977) has summarized work on rocks of the Niigata, Japan area by Iijima and Utada (1972). They investigated replacement minerals of rhyolitic pyroclastic rocks deposited in a marine environment, that have been subjected to temperature environments of 41-49^{o}C and higher (Fig. 3-10). Authigenic minerals in the temperature range of the Boise geothermal system (72^{o}C) should be montmorillinite, low-crystoballite, and alkali-clinoptilolite. At slightly higher temperatures mordenite, Ca-clinoptilolite, quartz, and celadonite should appear. The main difference in chemical environment from the Boise Geothermal System is that the formation water of the subsurface Niigata rocks may be connate water originally derived from sea water, and of different chemistry than the relativily fresh Na-HCO_3 water of the Boise wells.

Both zeolite and clay minerals can form from a volcanic-glass parent material in a variety of geologic settings, not necessarily associated with hydrothermal conditions. They form rapidly in alkaline (pH ~9.5), saline lakes, and more slowly in marine sedimentary environments. They also form in alkaline soil

Approximate depth limit of zone (km)	2	2.5	3.7	4.6	n.d.
Approximate upper temperature limit of zone (^{o}C)	41-49^{o}	55-59^{o}	64-91^{o}	120-124^{o}	n.d.

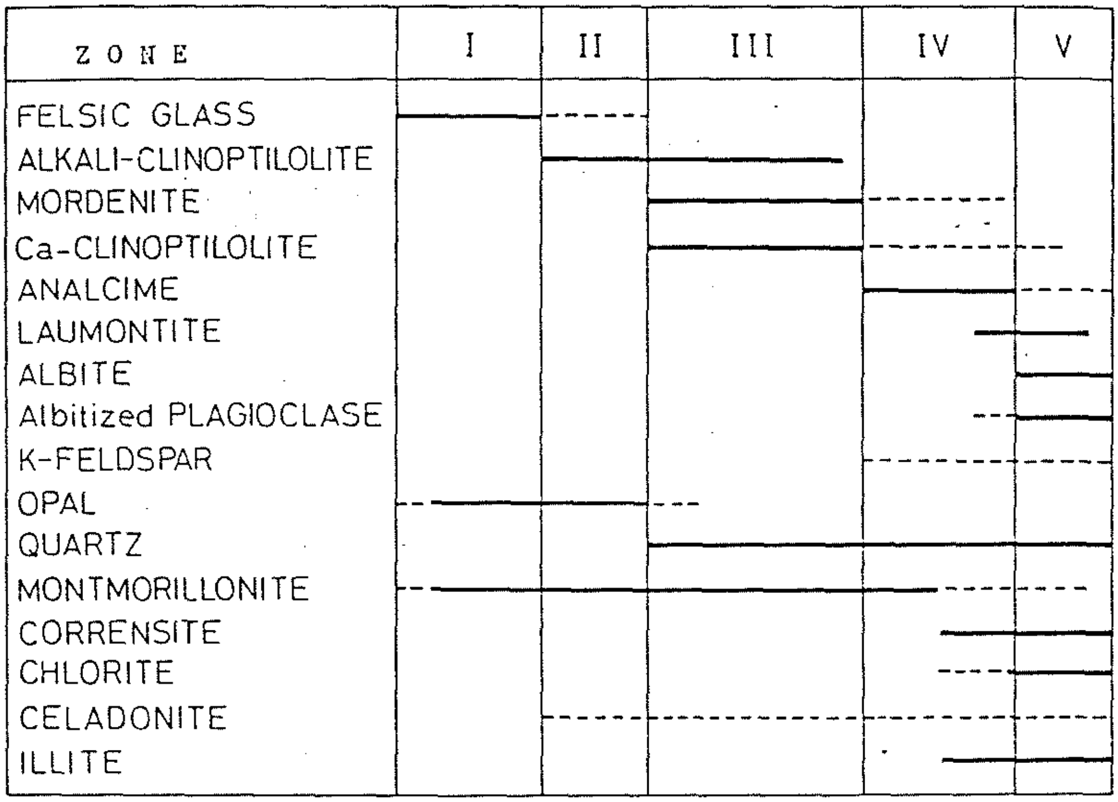

FIGURE 3-10. Stability ranges of authigenic minerals in volcaniclastic sediment and volcanic rocks of the Niigata (Japan) oil field (from Boles, 1977, after Iijima and Utada, 1972).

zones in some semi-arid regions. Thick sequences of volcaniclastic deposits alter to zeolite under normal groundwater flow-systems, and also undergo burial diagenesis in their own connate waters at very moderate temperatures (Hay, 1977). Temperature probably hastens the alteration, but the chemistry of pore water, initial glass chemistry, and time are clearly a factor (Boyle, 1977). Reaction kinetics in various geochemical settings seem to be an important determinant on the nature of the clay-mineral and zeolite assemblages. Boyle (1977) states that members of the heulandite and analcime groups can form at essentially standard temperature and pressure conditions. However, formation of laumontite and wairakite require elevated temperatures.

Holler and Wirsching (1978) carried out experimental alteration of volcanic glass at 150^{o} to $250^{o}C$. In open system experiments, water of fixed composition was percolated through the volcanic glass. In this system, the alteration rate is higher than in a closed system. The continuous addition of fresh solution favors fast dissolution of the glass. As a consequence, alteration minerals form rapidly. In strongly alkaline solutions, the addition of Na^{+} favors zeolite formation. If solutions are acidic to weakly alkaline, surplus alkalies are removed and clay minerals are formed. Resulting solutions also differed due to S_iO_2 activity related to initial composition of the glass. Mordenite, for instance, only formed in conditions of high S_iO_2 activity in their experiments and would be expected with rhyolite glass.

Because Na^+ is overwhelmingly the dominant cation in Boise geothermal water, formation of the sodium zeolite, stilbite is clearly favored. If heulendite is indeed present, it is also likely to be a Na-Heulandite. The clay mineral is expected to be a Na-montmorillinite. Formation of montmorillinite clay may be characteristic of the relatively low alkalinity of the water. The formation of these materials as veins in fissures clearly shows that <u>they are precipitates and not just replacement minerals</u> of the glass.

Significance of Iron Pyrite

Iron pyrite occurs in cuttings from geothermal wells from deeper than a few hundred feet. Iron pyrite is very insoluble in a reducing environment, and can be in equilibrium with water with small concentrations of sulfide species and ferrous iron. Boise geothermal water has not been analyzed for iron or sulfide ion species (H_2S, HS^{-1}, and S^{-2}), but comparison with geothermal waters elsewhere suggests HS^{-1} might be of the order of 0.1 mg/l, and iron of the order of 0.4 mg/l. Truesdell and Jones (1974) give the precipitation-solubility reaction at 25°C as:

$$FeS_2 + 2H^+ + 2e^- = Fe^{+2} + 2HS^- \qquad \log K = -18.479$$

and an expression for the equilibrium constant is

$$K = [Fe^{+2}]\,[HS^{-1}]^2 \,/\, [FeS_2]_s[H^+]^2\,[e^-]^2$$

Because HS^-, pH, and oxidation potential are all squared in this expression, these three components have a stronger influence on pyrite stability than the activity of ferric iron which is to the first power. Because of the great difference in oxidation potential of surface water in equilibrium with the atmosphere, and deeply percolating ground water, the greatest effect on the stability of pyrite is the oxidation potential of the water. At $25^{o}C$, the boundary between the oxidation of sulfide species to sulfate, or pyrite to hematite is at an oxidation potential of about 10^{-72} (expressed as partial pressure of oxygen in atmospheres). Therefore the introduction of even a small amount of oxygenated groundwater should destroy pyrite. The effect of temperature is to raise, the oxidation potential at which pyrite is oxidized, so that at $100^{o}C$, the oxidation boundary is about 10^{-52} (Barnes and Kullerud, 1961). Barnes and Kullerud also state the intuitive predudice that increasing temperature increases oxidation is a reflection of the rate of reaction and not an effect on equilbrium. Since none of the constituents in the the pyrite-water equilibrium have been measured in the geothermal waters, one can just speculate that the waters are in equilibrium with pyrite, and that the depth to first pyrite is related to the depth of percolation of oxygenated water.

Significance of silica

The silica concentrations in the Boise geothermal waters are

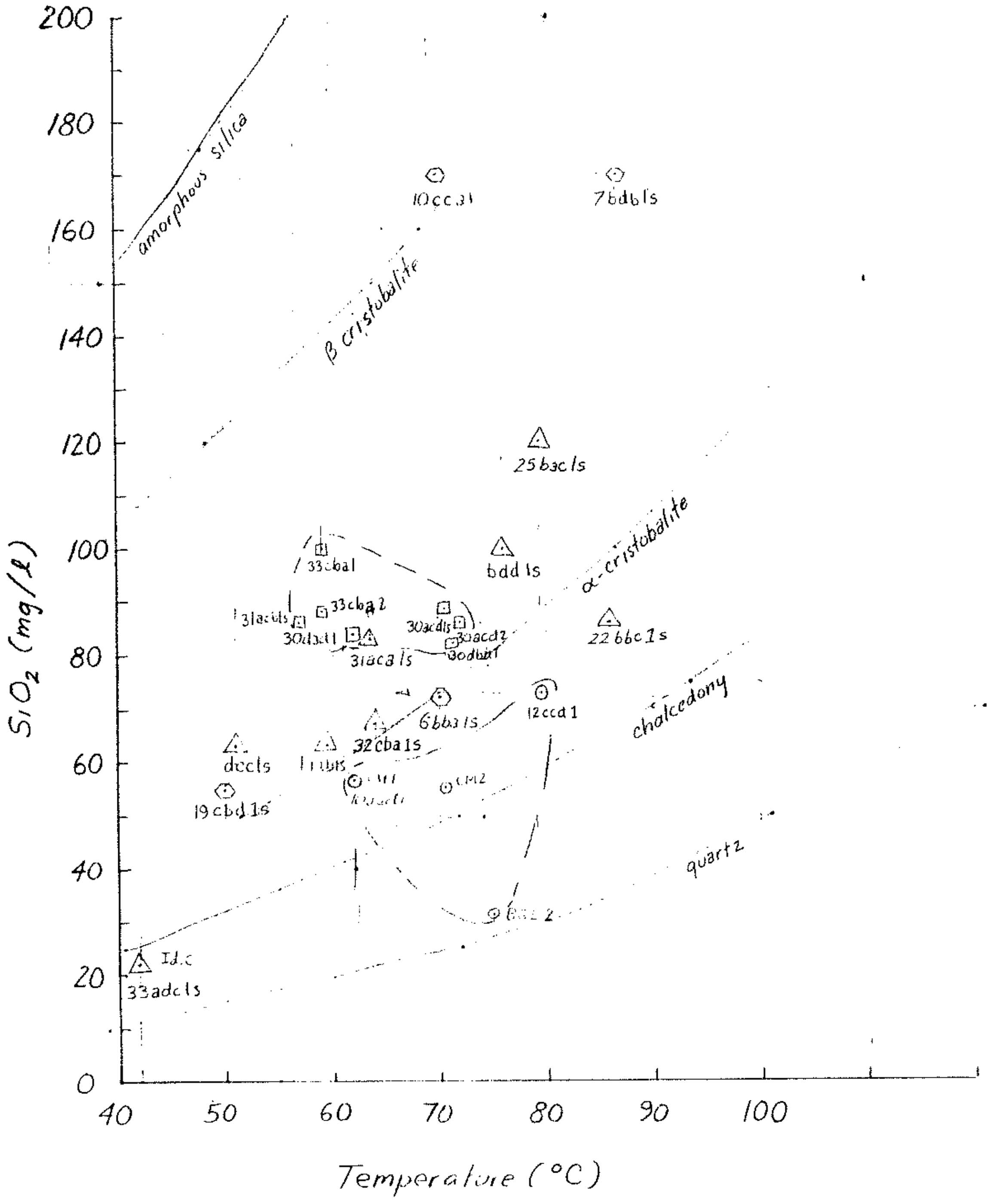

FIGURE 3-11. Silica concentration vs. temperature for selected geothermal waters compared to the Boise geothermal waters. Lines represent temperature-silica-solubility relationships for various silica minerals as given by Truesdell (1984).

30 to 80 mg/l, which for 60^{o}C water indicates equilibrium with the precipitation of chalcelony (30 mg/l) to beta-crystobalite (80 mg/l) (see Fig. 3-11). Waters from the Banbury and batholith springs are theoretically saturated with respect to alpha-crystobalite. Only the basalt aquifer waters approach saturation with respect to beta-crystoballite or amorphous silica.

The sandstones in the vicinity of the major fault of the Boise system are strongly cemented with chalcedony. Cuttings from rhyolite in the aquifer are laced with veins of green and red jasper. It is questioned whether these relatively low silica concentrations in the Boise geothermal water have prevailed over a long period of time and entirely account for the silica cementation of exposed permeable sand units near major fault zones. Normally, abundant siliceous sinter deposition occurs about springs for which underground "reservoir" temperatures are in excess of 180^{o}C (Ellis and Mahon , 1977, p.110). These are boiling springs with geysers, and with silica concentration in water in excess of 250 ppm. The silica cementation of the sands exposed in the foothills occurs within about 1500 feet of major faults indicating that water cooled and precipitated much of its silica within the first 1500 feet of
lateral migration in the sand aquifers.

Stable isotope studies (^{18}O and ^{2}H)

Stable isotope studies of geothermal groundwater in southern Idaho by Young and Lewis (1982) and from the Boise system (Mayo and others, 1984) show that thermal waters are shifted about -20 $^{o}/oo$ in $\delta^{2}H$ and -2 $^{o}/oo$ in $\delta^{18}O$ from the meteoric water line. In other words, the thermal waters are isotopically light with respect to both hydrogen and oxygen in comparison with present-day rainwater and cold springs of shallow groundwater circulation. Since variations in hydrogen isotopes are attributed entirely to atmospheric evaporation and condensation processes, the different $\delta^{2}H$ levels imply that either the climatic condition at the time of recharge or the elevation of the place of recharge is different from that of the thermal spring or well. Young and Lewis (1982) removed this ambiguity of $\delta^{2}H$ interpretation by analyzing cold-spring waters from 7000 ft elevation. High elevation spring waters were very similar in $\delta^{2}H$ values to cold spring waters at lower elevation. Therefore, altitude of recharge area does not explain the shift of $\delta^{2}H$ value in thermal spring waters from the meteoric water line (Fig. 3-12). Rather, they suggest that the isotope shift must be a relict of precipitation in an earlier time of different climate.

It is not presently possible to identify some unique time in the late Quaternary in which precipitation in southern Idaho had a $\delta^{2}H$ of around -140 o/oo rather than -120 $^{o}/oo$ of modern precipitation. What little work has been done on climatic variation in $\delta^{2}H$ (Yapp and Epstein, 1977, Epstein and Yapp, 1976; Harmon and others, 1979) has yielded conflicting results

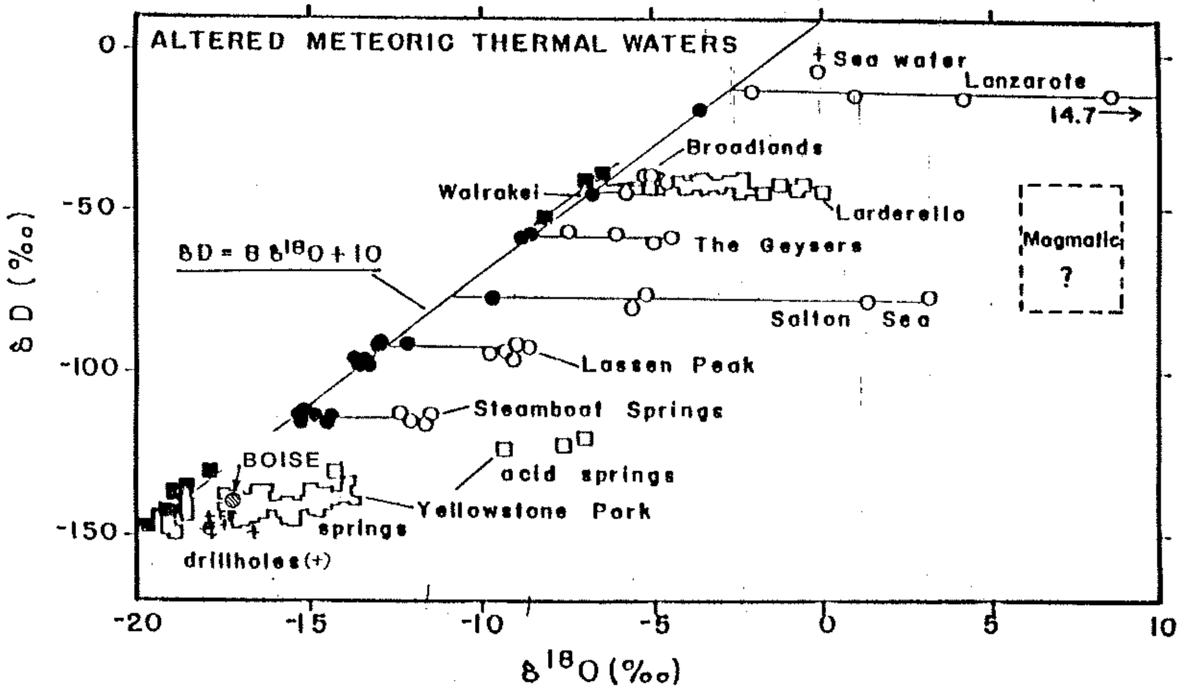

FIGURE 3-12. Plot of $\delta^{18}O$ vs $\delta^{2}H$ for geothermal waters from Truesdell (1984b). Isotopic composition of Boise geothermal waters from Mayo and others (1984) shown as cross-hatched circle within the group of values for thermal springs in Yellowstone Park. Other darkened symbols are meteoric waters, open symbols are thermal springs.

(Friedman, 1983). Some studies of groundwater in other regions that presumably originated as late Pleistocene precipitation have yielded a result of more negative δ^2H (isotopically light) values (Siegal and Mandle, 1984) similar to results from geothermal water of southern Idaho, while other studies have found no significant difference from modern day precipitation (Hanshaw and others, 1980). Therefore the hydrogen isotope data on the Boise water gives support to other data on the antiquity of the water, but it does not give a good constraint on its age.

Boise waters have a $\delta^{18}O$ value of -17.1 to -17.5 $^o/_{oo}$ which is isotopically lighter, more negative than the values from cold springs of the area which are -14 to -16 $^o/_{oo}$, a result identical to that obtained by Young and Lewis (1982) for geothermal waters of southern Idaho. The negative shift is unusual for geothermal water, because normally geothermal waters are shifted in a positive amount because of rock-water interaction. If there is chemical interaction of water with igneous rock, the $\delta^{18}O$ value is shifted to a heavier isotope value. On a plot of $\delta^{18}O$ vs. δ^2H plot (Fig. 3-12), values plotted for waters that have interacted with igneous rock minerals are shifted to the right of the meteoric water line. Meteoric recharge waters in the southern Idaho region range from $\delta^{18}O$ = -17.5 to -18.0 $^o/oo$. Most igneous rocks have $\delta^{18}O$ values of +4 to +13$^o/oo$ (Taylor, 1968, 1977). Waters from hot geothermal systems such as Steamboat Springs, Nevada and Yellowstone, Wyoming are typically shifted about +4 $^o/oo$ toward igneous rock values (Truesdell, 1984, p. 132-134). Lower temperature systems have lesser shifts.

However, $\delta^{18}O$ shifts from rock/water interaction are not simply temperature dependent. It has been speculated that in systems in which there is a large flux of recharge water relative to surface area of aquifer rock, the amount of water/rock interaction may be small, and hence the amount of $\delta^{18}O$ shift will be small. Such would be the case for a circulation system through relatively few, large fractures. Since both temperature of reacting water, and the flux of circulation can have similar effects on the value of $\delta^{18}O$ shift, the interpretation of oxygen isotope studies of geothermal waters can be ambiguous. The negative shift in oxygen isotope ratio can also be relict of water precipitated in an earlier climate. As pointed out by Mayo and others (1984, fig. 10), the Boise geothermal water falls within the scatter of worldwide points along the meteoric water line. Therefore, the isotopic composition of an earlier climate may have caused depletion of heavy isotopes relative to the composition of modern precipitation, and this effect may have overridden a small positive shift due to rock-water interaction.

Isotopic dating of Boise geothermal waters

Activities of radioisotopes, carbon-14 (radiocarbon) and hydrogen-3 (Tritium) have been measured in several wells in the Boise area, five springs in the Idaho batholith region, and in several hot wells in the Bruneau-Grandview area (Mayo and others, 1984; Young, 1985; and Young and Lewis, 1982, respectively).

These isotopic activities are useful in estimating the mean residence time that waters have been isolated as groundwater from the atmosphere.

Tritium

Natural tritium levels in rainwater are less than 10 T.U. (1.0 T.U. = 1 tritium unit = one atom of ^3H in 10^{18} atoms of common hydrogen). It is estimated that natural levels in groundwater should be lower than 2 to 4 T.U., but few measurements were made prior to the atmospheric testing of thermonuclear bombs. Thermonucluear bomb testing from 1953-1969 raised the tritium concentrations in northern (and presumsably worldwide) rainwater to several thousand tritium units. Since the USSR-US atmospheric testing moratorium in 1969, that level has declined and fluctuates within 30 to 100 tritium units. Half life of tritium is 12.3 years. Thus water with less than 5 to 10 T.U. is considered pre-1953 water (Freeze and Cherry, 1977, p. 136).

Tritium values of 3.4 and 1.6 $\pm$ 1.4 T.U. are obtained on waters from Capitol Mall wells 1 and 2, respectively (Mayo and others, 1984). Tritium values of 0.2 to 1.9 T.U. are reported for 7 thermal wells and springs south of the Snake River in southwestern Idaho (Young and Lewis, 1982). Tritium values in 13 thermal springs in the Idaho batholith region range from 0.1 to 4.1 _ 0.5 and average 0.8 T.U. (Young, 1985). Since most of this sampling was done about 1980, and all analysis are very low, we can conclude that no measurable amount of post-1953 meteoric water has mixed with these geothermal waters of southern Idaho.

Carbon-14 Dating of Groundwater

The carbon-14 activity of thermal groundwater gives a useful estimate of the maximum time that water has resided in the subsurface, totally isolated from the earth's atmosphere. Subsurface residence times have been estimated for Boise geothermal waters by Mayo and others (1984) and for batholith waters by Young (1985). The method depends upon an understanding of how dissolved carbon is introduced into the water. Ideally the main source of dissolved carbon would be from solution of carbon dioxide from the atmosphere and from soil gas; however these geothermal waters have acquired additional carbonate species during subsurface circulation. This is surprising, because these waters have not circulated in carbonate sedimentary rocks. Nevertheless, the geothermal waters have acquired additional subsurface carbon while circulating in igneous rocks. The added carbon with low or zero activity dilutes the amount of radiocarbon in the water sample and causes the calculated age of the waters to be too old. The trick is to estimate how much bicarbonate at atmospheric activity was emplaced in the recharging water, and how much bicarbonate at low or zero activity has been added along the circulation path. As will be shown later, the estimate of bicarbonate at atmospheric activity varies from 25 to 75 mg/l, and the final bicarbonate content of the Boise waters is about 111 mg/l, suggesting that at least 40 and as much as 90 mg/l of bicarbonate of low or zero radiocarbon activity has been added during circulation in the subsurface. This range of estimates leads to a correction of 3200 to 12,300 years that is subtracted from the calculated raw ages of Boise geothermal waters. The calculated raw ages from Mayo and other's (1984) data on water samples from four different wells range from 11,900 to 22,500 years.

One could consider the range of 11,900 to 22,500 years to be a maximum possible residence time for the Boise geothermal waters except for the problem of mixed waters which has not been dealt with by previous investigators. If cooler waters of shorter residence time have mixed with the deeply circulating waters, then the calculated raw ages will be too small. Of the four Boise samples, the Boise Warm Springs (BWSWD) well sample appears to be the least-mixed sample from the deep circulation system, because these waters have the the highest temperature and highest saturation with respect to fluorite when compared with the BGL and Capitol Mall wells. The BWSWD well water also has the lowest radiocarbon activity, and therefore the highest calculated raw age of 22,500 (Table 3-2). Applying the range of corrections above yields an adjusted residence time in the range of 10,200 to 19,300 years, for the BWSWD well. Younger ages of the other wells may be caused by mixing with younger groundwaters.

Corrections applied to radiocarbon-derived residence times.

Source of all known carbon-14 is from upper atmosphere interactions with cosmic radiation. The carbon-14 activity in groundwater cannot be directly converted to residence time, because the original carbon input has generally been diluted by input of less active or "dead" carbon from dissolution of carbonate minerals from rocks, veins or void fillings along the circulation path of the groundwater. This input of "dead carbon" clearly occurs in percolation through ancient carbonate sedimentary rocks, and surprisingly, it also occurs to some extent in percolation through crystalline igneous rocks (Fritz, 1982), but the source of added carbon from water interaction with igneous rocks is not always known. The addition of aged or dead carbon makes the calculated age of water appear older than the true age or true average subsurface residence time.

Infiltrating groundwater in the recharge area equilibrates with the CO_2 component of soil gas. It is generally found that soil gas has a carbon-14 activity that is the same as the atmosphere so that principles applied to dating of photosynthesized organic material also apply to groundwater recharge water (Rightmire and Hanshaw, 1973). However, as groundwater percolates, the dissolved carbonate species in the water may exchange with well-aged carbonate minerals of essentually zero carbon-14 activity. and it is clear that additional carbonate species are added to the water by dissolution of these minerals or from some unknown source. It is also possible that carbon-14 is removed by a net precipitation of carbonate from the water into veins and voids in the rock. Carbonate minerals exist in the deeper soil zones in semi-arid lands, in most sedimentary rock sequences, and as vein or pore-space fillings in crystalline igneous and metamorphic rocks, and in glassy volcanic rocks.

Many investigators attempt to correct measured carbon-14 activity by estimating the amount of exchange of the groundwater carbonate species derived from the atmosphere and the upper soil zone with dead carbonate minerals along the path. This would be almost impossible save for the distinctive stable carbon isotope ratios ($^{13}C/^{12}C$) of carbonate minerals relative to the ratios found in atmospheric carbon dioxide, carbon fixed by photosynthesis, and carbon dioxide of soil gas. Carbonate minerals are isotopically heavier (higher and more positive values of $\delta^{13}C$) as shown in Figure 3-13. The differences, however, are not great; there is a considerable overlap.

The methods of correction of measured carbon-14 activity of groundwater to give a value that more nearly reflects the radioactive aging of the water are by no means precise. First, the $\delta^{13}C$ of the soil-gas source has a considerable range. Second, the $\delta^{13}C$ of the mineral carbonate in the subsurface has an even greater range (Figure 3-13). For instance, Fritz (1982) finds that calcite from cores of the Stripa granite in the Swedish experimental underground waste repository site has $\delta^{13}C$ ranging from -15 to +15 o/oo, and overlaps the range of both the soil gas and the mineral carbonate. The third difficulty results from isotopic fractionation that is know to be associated with dissolution and reprecipitation of calcite. This third uncertainty has been theoretically modeled by Wigley and others (1978), and Fontes and Garner (1979), but in view of the uncertainties in the stable isotope composition of the dissolved

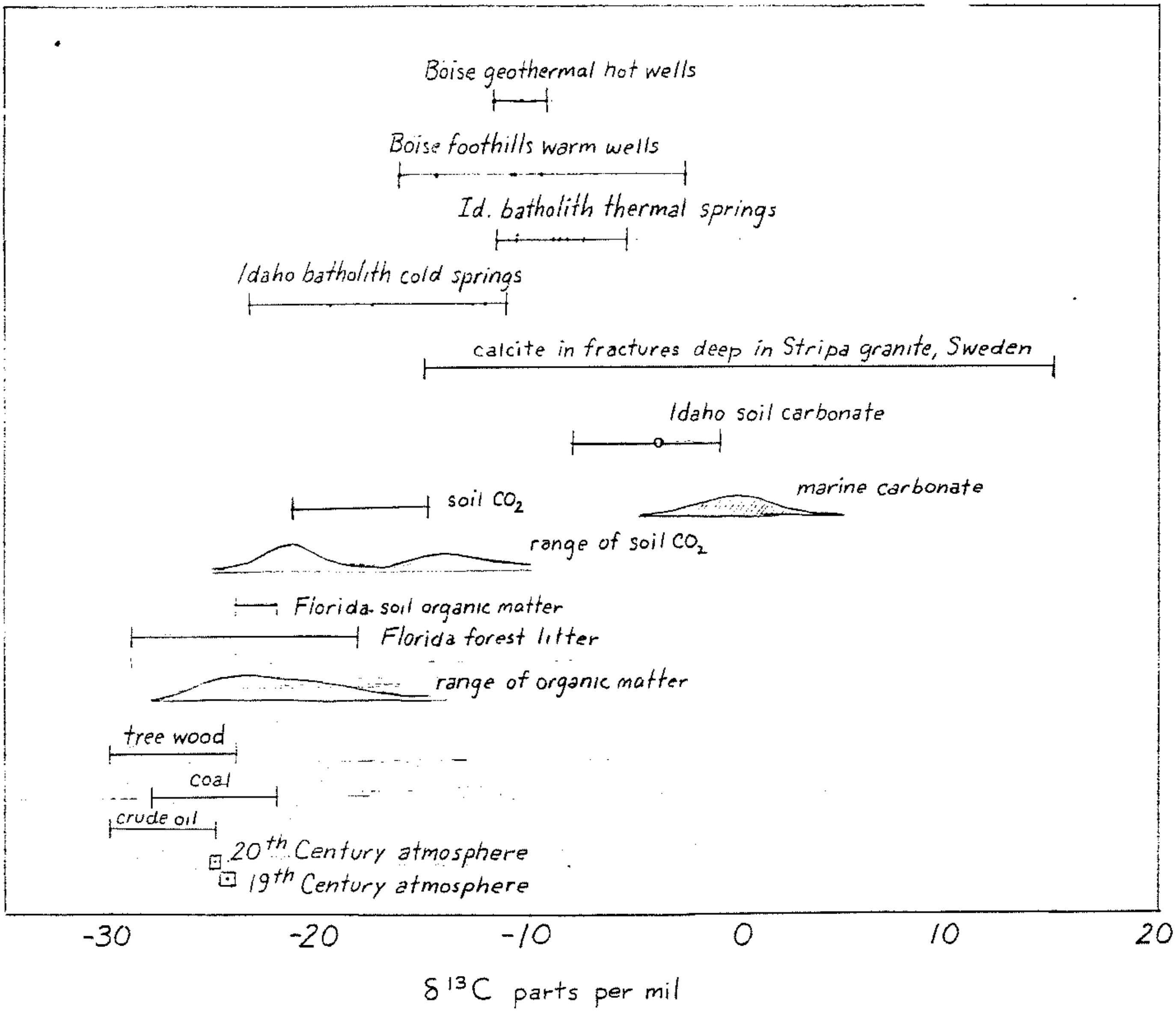

FIGURE 3-13. Diagram showing ranges of $\delta^{13}C$ for natural materials. Data are from Degens (1969), Rightmire and Hanshaw (1973), Fritz (1982), Young (1985), and Mayo and others (1984). Purpose of diagram is to show the the relatively wide ranges of $\delta^{13}C$ in subsurface-vein carbonate. Lack of knowledge of the stable-carbon isotopic composition of carbonates in the subsurface indicates that $\delta^{13}C$ cannot be used to correct carbon-14 activities to obtain a groundwater-residence time.

input carbonate species, and the mineral carbonates at depth, these theoretical calculations using $\delta^{13}C$ as an index of "dead carbonate" input to the system do not add much precision to the correction of measured residence time.

Mayo and others (1984) use a model involving carbonate reaction stoichiometry and values of δC^{13} from the geothermal water to calculate a correction for the residence times. They assumed the δC^{13} of the carbonate added in the subsurface to be zero. In Figure 3-14 and later discussion it is shown that δC^{13} of the subsurface additions must be in the range of -7.5 to -11.5 o/oo. Therefore, the method used by Mayo and others (1984) to correct their radiocarbon activity may not be valid.

Young (1985) uses a different approach in his study of the thermal springs of the Idaho batholith region. His method will be developed in the following discussion, and applied to Mayo and others' (1984) data so that residence times can be put on a comparable basis.

Young (1985) corrects carbon-14 activities using a method that does not rely on measured $\delta^{13}C$ values. He assumes that dead carbonate species have been added to the waters. The amount of dead carbon is measured simply as the increase in total carbonate species from the original infiltrating recharge water to the final carbonate species in the thermal spring water.
It is assumed that the bicarbonate value in cold-spring water is a good measure of the original carbonate. Thermal spring waters are considerably higher in total dissolved carbonate (HCO_3^{-1} and CO_3^{-2}). Young assumed that the apparent actvity measured in the

thermal water has been diluted by the added "dead carbonate". To correct for this added amount of "dead carbonate", the following calculation is derived after Young (1985).

The total carbonate species in the sample can be assumed to be the amount of the common stable isotope, carbon-12, in the original liter of water, since the amount of carbon-13 and carbon-14 are negligeable relative to carbon-12. In the original 1 liter of ground water:

$^{14}C_o$ = concentration of carbon-14 radioisotope at time zero

$^{12}C_o$ = concentration of total carbon at time zero.

After the liter of groundwater has passed through the circulation system and is sampled at the well or spring,

$^{14}C_t$ = sampled concentraton of the carbon-14 isotope. Carbon-14 is reduced by radioactive decay only (in this model, none is lost by exchange with carbonate minerals at depth).

$^{12}C_t$ = sampled concentration of the total carbonate species. This concentration is made up of the original total carbonate ($^{12}C_o$) plus mineral carbonate dissolved during flow ($^{12}C_a$).

$^{12}C_a$ = amount of subsurface mineral carbonate dissolved into one liter of subsurface water.

For the thermal spring or well, the measured ratio of carbon-14 to total carbon is $^{14}C_t/^{12}C_o$. An age calculated from this ratio will be too old because $^{12}C_t$ has an acquired component equal to $^{12}C_a$, so that the ratio appears to be leaner in carbon-14.

The correct value to use for age computation is $^{14}C_t/^{12}C_o$ but this ratio is not directly measurable.

Equation for the correct age is:

$$^{14}C_t/^{12}C_o = (^{14}C/^{12}C_o) \exp(-Lt)$$

or

$$\exp(-Lt) = (^{14}C_t/^{12}C_o)/(^{14}C^o/^{12}C_o)$$

where L = [ln 2]/[ln (half life)] = 0.693/5730

= 1/8266 years

This equation is algebraically manipulated by dividing the numerator and denominator of the right hand side by $^{12}C_t$, and rearranging the terms.

$$\exp(-L\,t) = [(^{14}C_t/^{12}C_t)/(^{14}C_o/^{12}C_o)] * (^{12}C_t/^{12}C_o)$$

The part of this equation without the factor $^{12}C_t/^{12}C_o$ is

$$\exp(-L\,t) = (^{14}C_t/^{12}C_t)/(^{14}C_o/^{12}C_o)$$

and is the equation for the unadjusted age, calculated as if no additional carbon-12 had been added to the carbonate system in the water. The factor $^{12}C_t/^{12}C_o$ is the correcton factor, and is simply the ratio of the total carbonate species of the input water to the total carbonate species concentration in the output water. In order to calculate the residence time adjusted for "dead carbon" dissolution, one uses the following equation.

$$-Lt = \ln\left[({}^{12}C_t/{}^{12}C_o)({}^{14}C_t/{}^{12}C_t)/({}^{14}C_o/{}^{12}C_o)\right]$$

The equation can also be viewed as

$$t = -(1/L)(pmc) - (1/L)\ln({}^{12}C_t/{}^{12}C_o)$$

adjusted age = unadjusted age - correction

where pmc = per cent modern carbon activity

$$= ({}^{14}C_t/{}^{12}C_t)/({}^{14}C_o/{}^{12}C_o)$$

In Young's (1985) work, the value of ${}^{12}C_t/{}^{12}C_o$ was obtained by taking a ratio of total carbonate species in thermal springs to the total carbonate species in a nearby, non-thermal, spring. A typical ratio is

$$(59 \text{ mg/l})/(23 \text{ mg/l}) = 2.60$$

resulting in an age correction of -7,900 years applied to an age of 19,000 years, which gives an adjusted age of 11,100 years

If Mayo and others's (1984) data on Boise waters are to be corrected so that they are compatible with Young's data on thermal springs in the batholith area, we must make a somewhat arbitrary choice of the carbonate species content of the water that first became isolated from soil gas (i.e. no further carbon-14 input). Young (1985) assumed values of nearby cold-water springs to correct the ages of the thermal waters in the batholith region. In order to visualize the possible choices, the total carbonate species analysis of non-thermal springs in the batholith (from work of Young, 1985) are plotted in histogram form (Fig. 3-15). The histogram shows three groupings (modes) of the data. One mode from 0 - 30 mg/l, another from 50 - 90 mg/l, and a third mode from 100 - 130 mg/l. The residence-time correction is related to the number chosen from this histogram, since the correction is proportional to the natural logarithm of the ratio of concentration in the thermal water to assumed or measured concentration of carbonate species in the original recharge waters with atmospheric radiocarbon activity. The histgram indicates that if the bicarbonate content of recharge water is similar to batholith cold springs waters, it could be from 25 to 75 mg/l. The unadjusted age of the BWSWD well water is 22,500 years (Capitol Mall no. 2 well). Total carbonate species in this well water is 111 mg/l, and this value is similar to other wells in the group. If we aarbitrarily choose an initial value of $^{12}C_o$ of 25 mg/l, then the correction (in years) applied to radiocarbon residence times is given by

$$(\text{half-life}/\ \ln 2)\ \ln(\text{total carbonate species}/\ 25\ \text{mg/l})$$

$$= (5730/0.693)\ \ln(111\ mg/l\ /\ 25\ mg/l)$$

$$= 8266\ \ln\ (111\ mg/l\ /\ 25\ mg/l).$$

$$= 12{,}300\ years$$

If, instead, we arbitrarily choose 75 mg/l for the input carbonate species, then the correction is 3,250 years.

Plotting Young's (1985) data in this histogram form (Fig. 3-15) shows a need for determining the radiocarbon activity of each of these modes of bicarbonate content of cold spring waters, before using bicarbonate content as a basis for adjusting radiocarbon-derived residence times. It would also be useful to obtain radiocarbon activity of carbonate minerals in deep mines in the batholith, particularly in zones of substantial groundwater flows. Such studies could aid in our understanding of the dynamics of recharge in the batholith and in the Boise system.

One other set of data has been plotted that may give a better basis for choosing a bicarbonate content of the "recharge water". The $\delta^{13}C$ values determined by Young (1985) and Mayo and others (1984) have been plotted versus the bicarbonate content of the waters (Fig. 3-14). The plot shows a roughly linear increase of bicarbonate content with $\delta^{13}C$ until the bicarbonate content reaches about 35 mg/l. For bicarbonate contents above this value the $\delta^{13}C$ values are all between -12.5

TABLE 3-2. ISOTOPIC AND OTHER DATA ON BOISE GEOTHERMAL WATER FROM MAYO AND OTHERS (1984) RECALCULATED.

Well Name	Temp. °C	pH	Fluor. solu. prod. $[Ca^{+2}][F^{-}]^2$ (mmoles/l)^3 (logarithm)	HCO_3 mg/l	$\delta^{13}C$ o/oo	$A^{14}C$ pmc	unadj. res. time in years	adj. res. time based on HCO_3]*	adj. res. time assuming $^{12}C_o$ is 0.5 dead carbon**
BWSWD	79	8.1	-9.9	110	9.3	6.6	22468	19302	17000
BGS#2	74	8.4	-9.2	111	9.5	11.1	18700	15460	12600
CM#1	65	8.5	-9.8	126	11.8	21	12900	8612	7100
CM#2	70	7.6	-10.1	114	10.5	23.7	11900	8440	6700

*assuming recharge water carried 75 mg/l of dissolved carbonate species from soil gas at atmospheric activity, and dead mineral carbonate was added to give the value of (HCO_3).

** assumption of Mayo and others (1984)

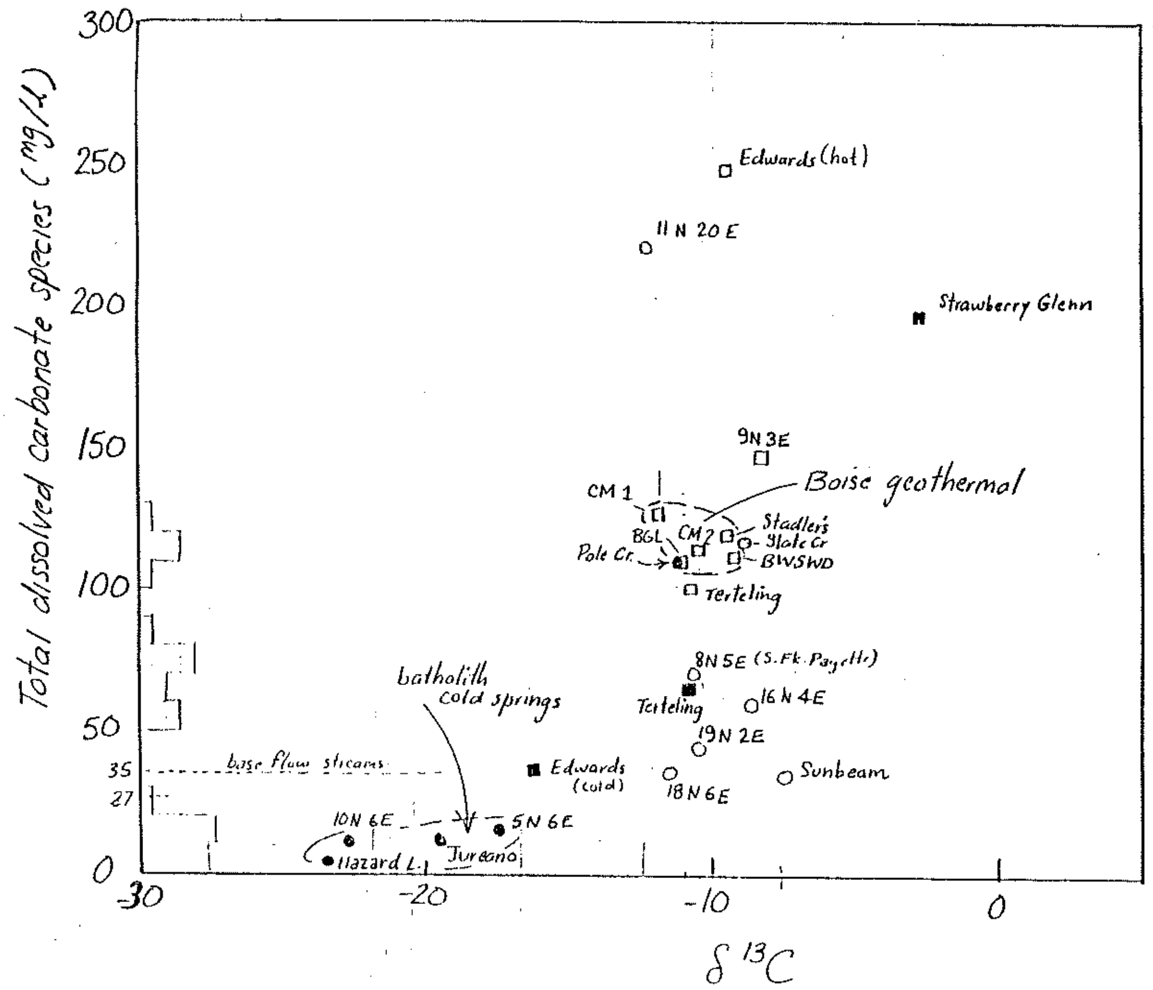

FIGURE 3-14. Plot of $\delta^{13}C$ vs. total dissolved carbonate species (bicarbonate ion conc.) of selected cold spring and warm spring waters of the Idaho batholith area and the Boise system. Data is from Mayo and others (1984) and Young (1985). The plot suggests that bicarbonate and $\delta^{13}C$ increase together as meteoric waters percolate through granitic rock and acquire bicarbonate from a subsurface carbon source. At a bicarbonate concentration of 25 to 75 mg/l the $\delta^{13}C$ reaches a near contstant value suggesting the bicarbonate acquired from subsurface source has a $\delta^{13}C$ between -12.5 and -7.5 o/oo, and not zero as is commonly assumed.

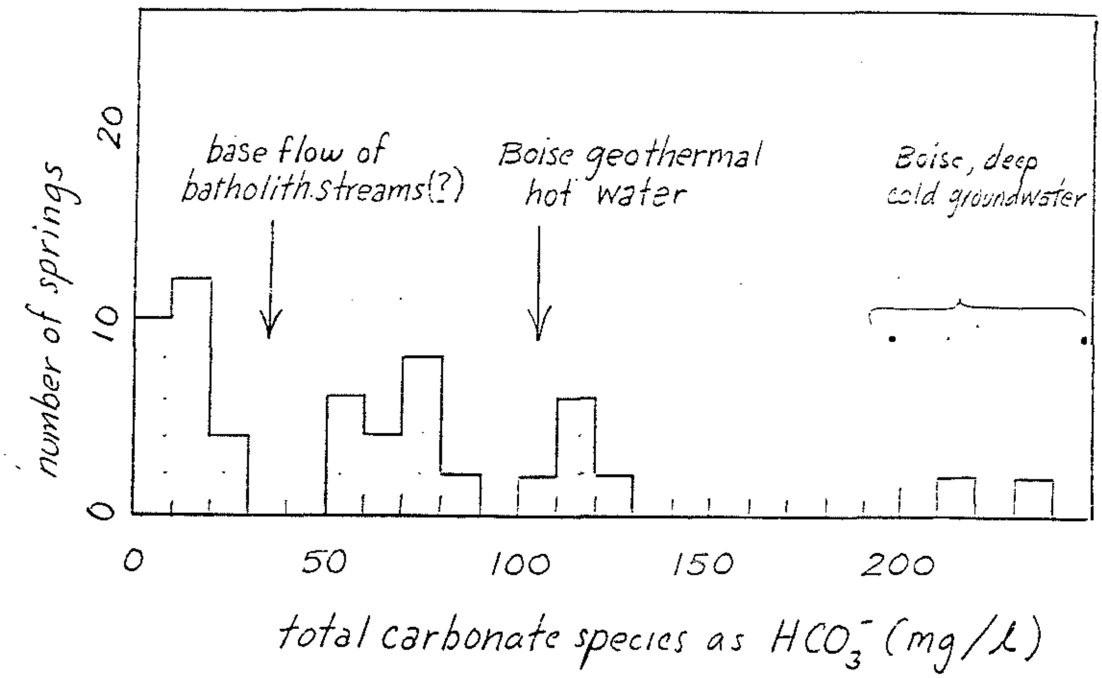

FIGURE 3-15. Histogram of analysis for total carbonate species (carbonate + bicarbonate ions) in cold spring waters of the Idaho batholith region. Data is from Young (1985). Purpose of the plot is to identify a concentration that is typical of recharge water when it first becomes isolated from the atmosphere and soil-gas input of radiocarbon.

and -7.5 o/oo. The suggestion from this plot is that waters less than about 25 mg/l have entirely atmospheric or soil-gas carbon and have the value for $\delta^{13}C$ of about -20 o/oo. These waters incorporate more bicarbonate from a subsurface source with $\delta^{13}C$ values that fall in the range of -12.5 to -7.5 o/oo. This plot of data supports a choice of 25 mg/l for the input bicarbonate content of "recharge water", and suggests that water with greater bicarbonate may have acquired the additional carbon from a subsurface source. The correction using 25 mg/l is 12,300 years, which applied to the unadjusted residence time of 22,500 years for the BWSWD water yields a corrected residence time of 10,200 years. However, if the same assumption of 25 mg/l is made for the input water for the CM2 well water, the calculated correction is 12,542 years which is greater than the unadjusted residence time. This discrepancy suggests that the CM2 and CM1 well water may be a mixture of waters with relatively long residence times with waters of shorter residence time.

Summary and Conclusions

Comparison of the hydrochemistry of geothermal waters of selected systems in southern Idaho shows that each system has several unique chemical characteristics not clearly related to different temperatures. These characteristics are peculiar to lithology of the circulation path or the aquifers. Waters from basalt aquifers are low in fluoride and high in sulfate compared to waters from granitic and rhyolitic aquifers. It is difficult to distinguish by chemistry waters from granitic regions from those that have resided in rhyolite aquifers, although lithium analyses in addition to the standard water analyses may be helpful in future work.

In the Boise system, subsurface residence times derived from radiocarbon activity, range from 6700 to 17,000 years in four analyses. These long subsurface residence times alone show that the concept of recharge is meaningless for groundwater development and management of such a system. The quantity of water that is important is the amount of natural discharge that can be captured by wells.

Data summarized in Table 3-2 indicates that BWSWD well water has the greatest carbon-14 residence time, the greatest temperature, and the lowest $\delta^{13}C$ compared with the other wells of the Boise system. The CM2 well has the shortest residence time, the lowest pH, and the lowest fluorite solubility product. These data suggest that the BWSWD water is the least mixed sample and is probably derived mostly from the deep circulation system, and that other wells draw some of their water from systems of shallower circulation.

It is recommended that future geochemical study be directed toward quantifying the mixing of deeply circulating water with the waters that have only circulated in the shallow stratified rocks. In order to distinguish the deeply-circulating-water component, future analyses should focus on constituents or properties of the deeply circulating water that might be conservative in the mixing process. Temperature, fluorite solubility products, and lithium are suggested, realizing that the calculations would have to allow for water-rock chemical equilibrium, heat exchange, and other processes in addition to simple dilution. Such a study could help to quantify the subsurface discharge of the deeply circulated water into the stratified rocks. Knowledge of the amount and the region of that discharge may help in design of future reinjection and production wells.

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